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**STRATIGRAPHY AND STRUCTURAL GEOLOGY
OF THE COWANSVILLE-SUTTON-MANSONVILLE
AREA IN THE APPALACHIANS OF
SOUTHERN QUEBEC**

M.J. Rickard



1991



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Cover description

Looking northeast toward Roundtop within the Sutton Mountains.

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STRATIGRAPHY AND STRUCTURAL GEOLOGY OF THE COWANSVILLE-SUTTON-MANSONVILLE AREA IN THE APPALACHIANS OF SOUTHERN QUEBEC

Abstract

This report presents a detailed account of the stratigraphy, and analyses the structure and metamorphic textures, of an area at the western edge of the Appalachian region in the southern part of the Eastern Townships of Quebec. Three broad northerly-trending structural zones with different stratigraphies are recognized: the western Oak Hill sequence, the central realm of Sutton schists and an eastern Mansonville sequence.

The Western Zone is interpreted to be a late Precambrian to Cambrian rifted continental margin, the Central Zone is a continental-slope flysch wedge and the Eastern Zone sequence represents oceanic deposits. Deformation commenced in the Early Ordovician with the obduction of serpentinite, and by the Middle Ordovician (Taconic Orogeny), arc collision caused strong folding and westward thrusting. A highly attenuated slide zone separates the western folds from a major anticline in the central schists. Stress perturbations across an embayment in the continental margin produced east-west cross folds in deeper parts and successively oblique fold trends. The peak of metamorphism accompanied the late development of the major Sutton Anticline, and its effects are superimposed on earlier structures to the west.

In the eastern part of the area, unconformable Upper Silurian slates were folded into a tight syncline by the Middle Devonian (Acadian) orogeny. Radiometric dating yields equivocal results for the age of metamorphism in the Central and Western zones; it may be Taconic or as young as Acadian.

Résumé

Le présent rapport fournit un compte rendu détaillé de la stratigraphie ainsi qu'une analyse de la structure et des textures métamorphiques d'un secteur situé à la limite ouest de la région des Appalaches, dans la partie sud des Cantons de l'Est, au Québec. On y observe la présence de trois larges zones structurales orientées vers le nord dotées de stratigraphies différentes: la séquence d'Oak Hill à l'ouest, le domaine central des schistes de Sutton et une séquence de Mansonville à l'est.

L'interprétation indique que la zone ouest est une marge continentale fissurée formée au cours de la période s'étendant du Précambrien supérieur au Cambrien, que la zone centrale est un prisme littoral à flysch de talus continental et que la séquence des zones de l'est correspond à des formations océaniques. La déformation s'est amorcée durant l'Ordovicien inférieur avec l'obduction de serpentinite; à l'Ordovicien moyen (orogénèse taconique), la collision d'arcs insulaires a provoqué un fort plissement et une poussée vers l'ouest. Une zone de glissement très atténuée sépare les plis à l'ouest d'un anticlinal important situé dans les schistes centraux. Des perturbations attribuables à des tensions de part en part d'une baie située dans la marge continentale ont produit des plissements transversaux d'orientation est-ouest dans les parties les plus profondes qui se sont progressivement transformés en des plis obliques. Le métamorphisme le plus intense a marqué la période terminale de formation de l'important anticlinal de Sutton; ses effets sont superposés à des structures plus anciennes à l'ouest.

Dans la partie est de la région, des schistes discordants du Silurien supérieur ont été plissés en un synclinal fermé par l'orogénèse du Dévonien moyen (Acadien). La datation par méthode radiochronologique donne des résultats équivoques quand il s'agit d'établir l'âge du métamorphisme des zones centre et ouest; ce métamorphisme peut remonter au Taconique ou à une époque encore plus récente, c'est-à-dire l'Acadien.

SUMMARY

The study area lies in the southernmost part of the western edge of the Canadian Appalachian region in the Eastern Townships of Quebec. Three broad northerly-trending structural zones with different stratigraphic sequences are recognized: the western Oak Hill sequence; the central realm of Sutton schists, which form a major anticlinal zone; and an eastern, essentially homoclinal Mansonville sequence.

The Sutton Anticline is part of the Green Mountain-Sutton Mountain Anticlinorium. To the south, in central and southern Vermont, Precambrian basement is exposed in the core of the anticlinorium. To the north, at Richmond, Quebec, there is a depression of the hinge so that the fold closes; it continues, somewhat offset, northeastward through the Notre Dame Mountains. To the west of the Sutton Anticline, Hadrynian and Cambrian rocks of the Oak Hill Group are folded in the Richford-Valcourt Syncline and the Enosburg Falls Anticlinorium. The western edge of the Oak Hill sequence structural zone is marked by steep to northwesterly overturned Middle and Upper Cambrian rocks along the line of the Hinesburg Thrust. Farther west, Cambrian rocks of the Stanbridge and Granby nappes lie above Middle Ordovician rocks. The frontal edge of the allochthon is "Logan's Line". On the eastern flank of the Sutton Anticline the Cambrian and Lower Ordovician rocks are of oceanic aspect with thick ultramafic and basic volcanic units of a serpentine belt extending northeastward to the Thetford Mines district. In the vicinity of Lake Memphremagog, Upper Silurian rocks occur in tight synclines. The results of stratigraphic and structural mapping are presented on a 1:50 000 scale geological map.

The Oak Hill sequence is composed of metabasalts, lesser metarhyolites and detrital-rich greywackes followed by dolomite. A disconformity marked by ferricrete separates these probable Upper Hadrynian sediments from Lower Cambrian phyllites and silty quartzites. The latter are overlain in turn by dolomite and quartzite and then by Middle and Upper Cambrian slates.

The repetition of the shallowing-upward quartzite-dolomite sequence above basic volcanic rocks in the Oak Hill Group may have resulted from the development of thermal-sag basins over the initially rifted continental margin.

The Sutton schists of the Central Zone comprise albitic and graphitic schists and gneisses with minor greenschist layers. The lithologies and chemistry suggest that the protolith represents a late Precambrian continental-slope flysch deposit on oceanic crust or thinned continental crust. The internal stratigraphy of the Sutton schists is interpreted to be a simple sequence, although small lenses of serpentinite mark possible early thrusts. There is, however, no evidence for large early nappes. The upper parts of the Sutton schists are probably correlatives of Oak Hill sediments;

SOMMAIRE

La région à l'étude est située dans la pointe la plus méridionale de la bordure ouest de la région des Appalaches canadiennes, dans les Cantons de l'Est, au Québec. On y reconnaît la présence de trois larges zones structurales orientées vers le nord et dotées de séquences stratigraphiques différentes: à l'ouest, la séquence d'Oak Hill, au centre le domaine des schistes de Sutton, qui forment une importante zone anticlinale, et à l'est, la séquence de Mansonville, essentiellement une structure homoclinale.

L'anticlinal de Sutton fait partie de l'anticlinorium de Green Mountain et de Sutton Mountain. Au sud, dans la partie centrale et dans la partie sud du Vermont, le socle précambrien est exposé au coeur de l'anticlinorium. Au nord, c'est-à-dire à Richmond au Québec, il existe une dépression de la charnière qui conduit à la fermeture du plissement; ce dernier se prolonge, mais d'une manière quelque peu décalée, vers le nord-est par les monts Notre-Dame. À l'ouest de l'anticlinal de Sutton, des roches de l'Hadrynien et du Cambrien du groupe d'Oak Hill sont plissées dans le synclinal de Richford et Valcourt et dans l'anticlinorium d'Enosburg Falls. La bordure ouest de la zone structurale de la séquence d'Oak Hill est marquée par des roches du Cambrien moyen et supérieur qui se dressent en pente raide ou sont renversées vers le nord-ouest le long de la ligne de faille inverse d'Hinesburg. Plus à l'ouest, les roches du Cambrien des nappes de Stanbridge et de Granby reposent sur des roches de l'Ordovicien moyen. Le front des terrains allochtones constitue la "ligne de Logan". Sur la bordure est de l'anticlinal de Sutton, les roches du Cambrien et de l'Ordovicien inférieur sont d'aspect océanique et comportent d'épaisses unités volcaniques de nature ultrabasique et basique d'une zone de serpentine qui se prolonge en direction nord-est jusque dans le district de Thetford Mines. Aux environs du lac Memphremagog, des roches du Silurien supérieur forment des synclinaux fermés. Les résultats des levés de cartographie structurale et stratigraphique sont présentés sur une carte géologique à l'échelle de 1/50 000.

La séquence d'Oak Hill se compose de metabasalts et de petites quantités de metarhyolites et de grauwwacks riches en matières détritiques et aussi de dolomie. Une discordance marquée par un conglomérat à ciment ferrugineux sépare ces sédiments, qui remontent probablement à l'Hadrynien supérieur, des phyllites et des quartzites silteux du Cambrien inférieur. À leur tour, ces derniers sont recouverts de dolomies et de quartzites, et par-dessus, d'ardoises du Cambrien moyen et du Cambrien supérieur.

La répétition de la séquence de quartzite et dolomie qui va en s'amenuisant vers le haut, au-dessus des roches volcaniques basiques du groupe d'Oak Hill, peut être le résultat du développement de bassins d'abaissement thermal situés au-dessus de la marge continentale qui était initialement faillée.

Les schistes de Sutton, dans la zone centrale, sont composés de schistes et de gneiss albitiques et graphitiques avec de petites couches de schistes verts. Les lithologies et la chimie laissent supposer que le protolithe représente un dépôt de flysch de talus continental datant du Précambrien supérieur et mis en place sur la croûte océanique ou la croûte continentale amincie. L'interprétation de la stratigraphie interne des schistes de Sutton est qu'il s'agit d'une séquence

there is a general eastward thickening of shale-silt facies but some Oak Hill formations are thinned by sedimentary and tectonic processes, especially at the contact with the Sutton schists.

The hypothesis that the Sutton schists represent a continental-rise prism implies that the schists cover the continent-ocean crust boundary or lie on thinned continental crust. This boundary is inferred to be located about 20 km west of the ophiolite belt, where it had previously been interpreted to be located. The boundary is marked by high gravity and magnetic anomalies under the Enosburg Falls Anticlinorium.

Middle Cambrian slates (Sweetsburg Formation) have been traced from the west to the eastern limb of the Sutton Anticline, where they are separated from a Cambrian to Middle Ordovician oceanic facies by a serpentinite belt that marks an obduction suture. These greywacke, shale and melange formations with thick metabasalts are unconformably overlain by Upper Silurian sediments.

During subsequent deformation the locations of the major structures were controlled by the paleogeography: the Enosburg Falls Anticlinorium developed more or less along the continent-edge upwarp geanticline; the Sutton Mountain Anticlinorium developed in the collapsing continental-rise prism under increasing metamorphic conditions; the obduction suture developed at the seaward edge of the continent-rise prism; and later large thrusts developed along the edge of carbonate banks to the west.

In the Western Zone, folds are generally upright and gently plunging in the Enosburg Falls Anticlinorium. A considerable variation of fold-axes orientations occurs in incompetent beds, as fold hinges have been drawn into the stretching lineation direction in slates. To the west, axial planes flatten to dip moderately eastward into the Hinesburg Thrust Zone. To the east in the Richford-Valcourt Syncline, Oak Hill rocks are strongly attenuated in the Brome Slide Zone and tight steep minor folds are common.

The Central Zone of Sutton schists is a simple major anticline of bedding schistosity, but early minor folds are east-west cross folds. These extend outward as steep reclined folds and are superimposed on the gently-plunging northeasterly-trending major folds. Late minor chevron folds on the Sutton Anticline trend more northerly and transect the axial trace of the major fold slightly. These folds and an associated crenulation cleavage are superimposed on all of the earlier structures to the west.

To the east, the cross folds become large near-vertical folds with a northerly-trending slaty cleavage, but the late Sutton Anticline structures die out eastward. At Lake Memphremagog, the Upper Silurian above the unconformity is folded into a gently-plunging syncline, but its axial-plane slaty cleavage does not penetrate the older rocks; the fold must have formed by shearing and tightening of basement folds.

simple, bien que de petites lentilles de serpentinite indiquent peut-être que des poussées très anciennes ont eu lieu. Cependant, rien n'indique la présence de grandes nappes anciennes. Les parties supérieures des schistes de Sutton sont probablement en corrélation avec les sédiments d'Oak Hill; il y a un épaississement général vers l'est du faciès de schistes et silt, mais certaines formations d'Oak Hill sont amincies par des processus sédimentaires et tectoniques, notamment au contact des schistes de Sutton.

L'hypothèse selon laquelle les schistes de Sutton représentent un prisme de glaciaire continental signifie que les schistes recouvrent la frontière entre la croûte continentale et la croûte océanique ou qu'ils reposent sur une croûte continentale amincie. On en déduit que cette frontière est située environ 20 km à l'ouest de la zone d'ophiolites, où l'on croyait qu'elle était située auparavant. La frontière est caractérisée par des anomalies gravimétriques et magnétiques importantes sous l'anticlinorium d'Enosburg Falls.

Des schistes ardoisiers du Cambrien moyen (formation de Sweetsburg) ont été retracés vers l'ouest jusqu'au flanc est de l'anticlinal de Sutton, où ils sont séparés d'un faciès océanique remontant du Cambrien à l'Ordovicien moyen par une zone de serpentinite qui indique la présence d'une suture d'obduction. Ces grauweekes, schistes ardoisiers et les formations mixtes avec d'épais metabasalts sont recouverts de façon discordante par des sédiments du Silurien supérieur.

Lors de déformations subséquentes, l'emplacement des principales structures a dépendu de la paléogéographie : l'anticlinorium d'Enosburg Falls s'est développé plus ou moins le long du géanticlinal à la limite continentale; l'anticlinorium de Sutton Mountain s'est développé dans le prisme littoral en effondrement sous le jeu du métamorphisme qui s'accentuait; la suture d'obduction s'est formée le long du prisme littoral, du côté de la mer; plus tard, des poussées importantes ont joué le long de la limite de bancs de roches carbonatées à l'ouest.

Dans la zone ouest, les plis sont généralement droits et plongent légèrement dans l'anticlinorium d'Enosburg Falls. L'orientation des plis varie considérablement dans les couches incompetentes, là où les charnières ont été entraînées dans la linéation d'allongement des schistes ardoisiers. À l'ouest, les plans axiaux s'aplatissent et plongent modérément vers l'est dans la zone de chevauchement d'Hinesburg. Vers l'est, dans le synclinal de Richford-Valcourt, des roches d'Oak Hill sont considérablement atténuées dans la zone de glissement de Brome et les microplis aigus et raides sont abondants.

La zone centrale des schistes de Sutton est un grand anticlinal simple à foliation parallèle à la stratification, mais les plis mineurs plus anciens sont des plis transversaux disposés d'est en ouest. Ces derniers s'étirent vers l'extérieur en plis raides et couchés et viennent se superposer aux plis majeurs qui plongent doucement en direction nord-est. Les plis mineurs en chevrons, assez récents, de l'anticlinal de Sutton sont davantage axés vers le nord et recoupent légèrement la trace axiale du pli majeur. Ces plis et leur schistosité de crénulation se superposent à toutes les structures plus anciennes qui sont situées à l'ouest.

À l'est, les plis transversaux deviennent de larges plis presque verticaux à clivage schisteux orientés vers le nord,

All minor structures have been analyzed in detail and the geometrical relationships presented on structural maps and stereograms.

Metamorphism was generally low but it increased to middle greenschist facies in the Sutton schists. Early segregations of quartz veinlets lie in the bedding schistosity and these are folded by early and late folds. The peak of metamorphism occurred during and after the late folding in the Sutton Anticline. Porphyroblasts of albite and rare garnet and chloritoid overprint the late crenulation cleavage; biotite is rarely developed. Petrofabric studies show that in spite of late porphyroblast growth, quartz orientations reflect both early and late folds. Outside the Sutton schists, the metamorphic effects are mild, but new growths of tourmaline and chlorite extend to the edges of the area.

The deformational history is related to current plate tectonic models. From the late Precambrian until the Early Ordovician the paleogeography was that of a miogeoclinal or continental-rise prism. The continent edge probably originated as a rift fracture in older basement, and transform offsets located re-entrants in the margin in which thick clastic wedges accumulated during the late Precambrian and Early Cambrian. A continent-edge upwarp at the fractured margin gave rise to a geanticlinal structure but by the Late Cambrian a broad sag basin had developed across the continental-rise prism and continental edge.

Major deformation commenced in the Early Ordovician when obduction of serpentinite occurred and the Atlantic-type margin was converted into a fringing island-arc system over an east-dipping subduction zone in the Middle Ordovician. These initial movements were accompanied by a rise of temperature in the thick continental-rise prism, and some thrusting involving oceanic basement slices (now serpentinite lenses) probably occurred. Quartz segregated into a schistosity developing parallel to bedding, and as compression increased some thrust telescoping may have occurred. Ordovician strata below the obduction zone probably piled up over the continental-rise prism and the dehydration of the over-thickened pile generated sufficient fluid pressure to initiate nappe sliding. As nappes slid or were pushed westward to form the Stanbridge and Granby nappes and Taconic nappes to the south, the continental-rise prism began to dome upward as it was compressed against the edge of the continental crust.

A continuation of compression in the Middle to Late Ordovician caused the Oak Hill strata on the continental shelf to crumple and slide westward on the Hinesburg Thrust as a para-autochthonous sheet, and a fanning system of folds developed to form the Enosburg Falls Anticlinorium. The fanning axial planes and the associated eastward thinning of the Oak Hill strata suggest a concave upward form for the Hinesburg Thrust, and a gravity sliding off the rising metamorphic core zone situated in the continental-rise prism. The Brome Slide Zone is interpreted to be a

mais les structures anticlinales de Sutton plus récentes se terminent en biseau à l'est. Au lac Memphrémagog, les terrains du Silurien supérieur au-dessus de la discordance sont plissés en un synclinal à plongement léger, mais leur clivage schisteux dans le plan axial ne pénètre pas dans les roches plus anciennes; le pli a dû se former par cisaillement et raidissement des plis du socle.

Toutes les structures secondaires ont été analysées en détail et leurs rapports géométriques ont été présentés sur des cartes structurales et des stéréogrammes.

Généralement, il y a eu peu de métamorphisme, mais qui s'est accru jusqu'à atteindre le faciès des schistes verts dans les schistes de Sutton. D'anciennes veinules de quartz qui ont migré selon la foliation parallèle à la stratification ont été déformées par des plis anciens et récents. Le métamorphisme a atteint un maximum durant et après le plissement récent dans l'anticlinal de Sutton. Des porphyroblastes d'albite et de rares grenats et chloritoïdes se superposent à la schistosité de clivage récente; il y a rarement formation de biotite. Les études de pétrographie structurale montrent que malgré la formation récente de porphyroblastes, les orientations de quartz sont un reflet de plis anciens comme de plis récents. Hors des schistes de Sutton, les effets métamorphiques sont légers, mais on observe que de la tourmaline et de la chlorite nouvellement formées se prolongent jusqu'à la limite du secteur.

L'évolution de la déformation correspond aux modèles courants de tectonique des plaques. Du Précambrien supérieur à l'Ordovicien inférieur, la paléogéographie était celle d'un miogéoclinal ou d'un biseau de glacis continental. Le bord du plateau continental tire probablement son origine d'un rift dans le socle ancien et d'indentations situées dans les zones de décalage créées par des failles transformantes au niveau de la marge où d'épaisses couches clastiques se sont accumulées au cours du Précambrien supérieur et du Cambrien inférieur. Un bombement de la marge à hauteur des fractures a conduit à la formation d'une structure géanticlinale, mais au Cambrien supérieur, un vaste bassin d'abaissement s'était formé sur le biseau de glacis continental et la marge du plateau continental.

La principale déformation a commencé durant l'Ordovicien inférieur, alors qu'il y a eu obduction de la serpentinite et que la marge passive a été convertie en un réseau d'arcs insulaires de bordure au-dessus d'une zone de subduction qui a plongé vers l'est durant l'Ordovicien moyen. Ces mouvements initiaux se sont accompagnés d'un réchauffement de l'épais biseau de glacis continental ainsi que, tout probablement de poussées qui ont impliqué des couches du socle océanique (maintenant devenues des lentilles de serpentinite). Le quartz a migré selon une foliation en développement parallèle aux couches et à mesure que la compression augmentait, il a pu se produire du télescopage de poussée. Les strates de l'Ordovicien situées en-dessous de la zone d'obduction se sont probablement empilées au-dessus du biseau de glacis continental et la déshydratation de cette amas surépaissi a créé assez de pression hydraulique pour amorcer le glissement des nappes. À mesure que ces dernières glissaient ou qu'elles étaient poussées vers l'ouest pour devenir les nappes de Stanbridge et de Granby, et devenir les nappes taconiques vers le sud, le biseau de glacis continental a commencé à se bomber à mesure qu'il se trouvait comprimé contre la limite de l'écorce continentale.

complex strain discontinuity located above the crustal hinge zone that during its subsequent development became folded by the Richford-Valcourt Syncline. The metamorphic boundary of the albite-porphyroblast zone coincides with the Brome Slide Zone in the west and the serpentinite belt in the east, suggesting that although the peak of metamorphism was later, the developing metamorphic belt controlled the location of structural breaks.

In the continental-rise prism itself, uplift continued and longitudinal shear on the bedding schistosity caused minor east-west reclined folds to develop across the regional trend. These extended upward to impinge on the lower part of the Oak Hill sequence in the Richford-Valcourt Syncline. The shearing movements were probably caused by differential lateral adjustments as the collapsing continental-rise prism was bent into the Quebec re-entrant in the Precambrian Shield, and by differential flow off rising metamorphic highs. The series of reclined to vertical folds in the Eastern Zone probably owe their spectacular development to their location in the concave core of the re-entrant.

The Taconic Orogen, thus formed, was quickly eroded and overlapped by shale-carbonate sedimentation in the Late Silurian to Early Devonian as the focus of orogenic activity shifted eastward. Final closure of the Iapetus Ocean occurred diachronously in the Acadian Orogeny, from Early to Late Devonian times as a continent-continent collision. There was only mild folding of the sediments covering the Taconic Orogen.

West of the serpentinite belt, deformation was terminated by a tightening phase in which the Sutton Anticline took on its final form. Thermal metamorphism reached its peak during this deformation. The obliquity of each successive phase of folding is related to compressive adjustments against the re-entrant in the continental margin.

Attempts to date the deformation phases radiometrically yielded equivocal results, as dates as old as 440 Ma and as young as 367 Ma are preserved in the Sutton schists.

La compression soutenue jusqu'à l'Ordovicien moyen et supérieur a provoqué le froissement des strates d'Oak Hill sur le plateau continental et leur glissement vers l'ouest sur la surface de charriage d'Hinesburg sous forme d'une nappe para-autochtone, et en un réseau de faisceaux de plis en éventail qui en se développant a formé l'anticlinorium d'Hinesburg Falls. Les plans axiaux en éventail et l'aminçissement associé vers l'est des strates d'Oak Hill semblent indiquer que la zone de charriage d'Hinesburg accuserait forme concave ascendante et qu'il y aurait eu un glissement gravitaire le long du noyau métamorphique en soulèvement situé au sein du biseau de glaciis continental. L'interprétation donnée à la zone de glissement de Brome est celle d'une discontinuité complexe résultant de contraintes et située au-dessus de la zone charnière du socle qui a été plissée par le synclinal de Richford et Valcourt au cours de son évolution subséquente. La frontière métamorphique de la zone à albite et porphyroblaste coïncide avec la zone de glissement de Brome à l'ouest et avec la zone de serpentinite à l'est; ce phénomène mène à penser que même si le maximum de métamorphisme a eu lieu par après, l'emplacement des ruptures structurales a été déterminé par le bombement géant-clinal métamorphique en formation.

Dans le biseau de glaciis continental lui-même, le soulèvement s'est poursuivi et un cisaillement longitudinal exercé sur la foliation parallèle à la stratification a conduit à l'apparition de petits plis couchés et orientés d'est en ouest, en travers de la direction générale dans la région. Ces plis se sont prolongés vers le haut et ont empiété sur la partie inférieure de la séquence d'Oak Hill dans le synclinal de Richford et Valcourt. Les mouvements de cisaillement ont probablement été causés par des ajustements latéraux différentiels à mesure que le biseau de glaciis continental qui s'effondrait était replié dans l'indentation du Québec au sein du Bouclier précambrien ainsi que par un flux différentiel sur les sommets des structures métamorphiques en soulèvement. La série de plis couchés à verticaux de la zone est doit probablement son évolution spectaculaire à son emplacement au sein du noyau concave de l'indentation.

L'orogène taconique ainsi formé s'est rapidement érodé et il s'y est superposé une sédimentation de roches carbonatées et de schistes ardoisiers durant le Silurien supérieur et jusqu'au Dévonien inférieur, alors que le foyer d'activité de l'orogène se déplaçait vers l'est. La fermeture finale de l'océan Iapet s'est produite diachroniquement pendant l'orogénèse acadienne, soit du Dévonien inférieur au Dévonien supérieur, et a pris la forme d'une collision entre deux continents. Il y a eu seulement un plissement léger des sédiments couvrant l'orogène taconique.

À l'ouest de la zone de serpentinite, la déformation s'est terminée par une phase de plissement aigu durant laquelle l'anticlinal de Sutton a pris sa forme définitive. Le thermométamorphisme a atteint un maximum durant cette déformation. L'obliquité de chaque phase successive de plissement est fonction des ajustements par compression exercés contre l'indentation à hauteur de la marge continentale.

Les tentatives de datation par méthode radiochronologique des phases de déformation ont donné des résultats équivoques : certains des schistes de Sutton ont été datés à 440 millions d'années et d'autres à 367 millions d'années.

INTRODUCTION

The area studied is situated at the western edge of the Appalachian mountain system about 80 km east-southeast of Montreal in the Eastern Townships of Quebec. It is covered by the Sutton (NTS 31 H/2) and western half of the Memphremagog (NTS 31 H/1) topographic maps and lies between latitudes 45°00' and 45°15' north, and longitudes 72°15' and 72°55' west.

Topography

Bedrock strongly controls the topographic features. The Sutton Mountains (highest point is Roundtop, 960 m) form a central belt of high terrain, coinciding with a major anticline of schists, that extends northerly from Massachusetts into Quebec through the Green Mountains of Vermont, and through the Sutton Mountains and Notre Dame Mountains of Quebec almost to the latitude of Quebec City. The Missisquoi River valley and Bolton Pass cut obliquely across the range along prominent zones of east-west folding.

On the western side of the Sutton Mountains, the Sutton River valley is bordered on the west by two parallel belts of hilly country, the highest points of which are Pinnacle Mountain (700 m) in the southeastern belt, and Oak Hill (290 m) in the northwestern belt. These are composed of erosion-resistant formations of the Oak Hill Group. East of the Sutton Mountains, and between the North Missisquoi River valley and the western side of Lake Memphremagog, are a series of ridges and isolated mountains (Pevee, 645 m; Sugarloaf, 650 m; Owl Head Mountain, 735 m) carved in the massive Bolton lavas.

Rock exposures are generally plentiful, although thick glacial drift covers some critical areas especially along the sides of the wider valleys. The glaciated terrain provides low rounded outcrops that are often unsuitable for measuring structural elements. Outcrops are becoming overgrown with shrubs as a result of the recent reduction in farming activity.

Access

Access is generally good, there being numerous roads and trails; some parts of the Sutton Mountains are difficult to traverse, however, because of dense growth of old timber or new growth in lumbered areas.

Regional geology

Major structural elements are depicted in Figure 1. The Sutton Anticline is part of the great Green Mountain-Sutton Mountain Anticlinorium in schists of Hadrynian

(?) and Cambrian age. To the south, in central and southern Vermont, Precambrian basement is exposed in the core of the anticlinorium. To the north, at Richmond, Quebec, there is a depression of the hinge so that the fold closes; it continues, somewhat offset, northeastward through the Notre Dame Mountains.

To the west of this anticline, Hadrynian and Cambrian rocks of the Oak Hill Group are folded in the Richford-Valcourt Syncline and Enosburg Falls Anticlinorium. The western edge of this zone is marked by steep to northwesterly overturned Middle and Upper Cambrian rocks along the line of the Hinesburg Thrust. Farther west, Cambrian rocks in the Stanbridge and Granby nappes lie above Middle Ordovician rocks which occupy the northward continuation of the St. Albans Synclinorium. The frontal edge of the allochthon is "Logan's Line", to the northwest of which is a zone of thrust-imblicated Ordovician strata adjacent to the platform sediments underlying the St. Lawrence Lowlands.

On the eastern flank of the Sutton Anticline the Cambrian and Lower Ordovician rocks are of oceanic aspect with thick ultramafic and basic volcanic units of the serpentinite belt extending northeastward to the Thetford Mines district. In the vicinity of Lake Memphremagog, Upper Silurian rocks occur in tight synclines and farther to the east extensive outcrops of Siluro-Devonian rocks form part of the Connecticut Valley-Gaspe Synclinorium.

Brome Mountain is a Cretaceous "Montegian" intrusion (Currie, 1976) and its geology is not dealt with in this report.

Method of study

Field work was carried out in the summers of 1957 through 1959 and for 3 weeks in 1976. The eastern part of the outcrop of the Oak Hill Group was mapped in detail (1:16 000 scale) aided by aerial photographs; the remainder of the area was mapped on a 1:25 000 scale. The flanks of the Sutton Mountains were traversed at about 1 km intervals, and the central part at about 2 km intervals. In the western part of the area from Cowansville to Frelighsburg, mapping concentrated on tracing the Dunham dolomite outcrop, aided by T.H. Clark's original field maps. P.H. Osberg kindly allowed the use of his unpublished manuscript map to complete the northeastern corner of the map area north of Bolton Pass. Field work concentrated on the measurement of structural elements (some 8000 were recorded); laboratory work consisted of structural analysis, petrographic examination of 200 selected samples and some petrofabric studies. Chemical analysis of ten rock samples was done to ascertain the effects of metamorphism and possible metasomatism.

The results are presented on a geological map and cross section, detailed structural map (Fig. 5) and a series of structural stereograms (Fig. 6) (all in pocket). Locations for specimens and figures are given by map references in Zone 18T of the 1000 m Universal Transverse Mercator grid.

Previous work

The anticlinal nature of the Sutton schists has been known from the earliest reconnaissance work of Logan (1863), Selwyn (1883) and Ellis (1894). The first detailed work was by Clark, Fairbairn and McGerrigle, mapping for the Geological Survey of Canada from 1923 to 1931. Clark (1934, 1936a) summarized the stratigraphy and outlined the structure of the area, but the map was never published until revised by Eakins (1964). The western edge of the area, from Frelighsburg to Cowansville (Charbonneau, 1975b, 1980), and part of the northeastern area around Bolton-Centre and Mount Sugarloaf (Lamothe, 1979, 1981), have been remapped.

Clark (1934) postulated a series of thrust slices in which folded lower Paleozoic geosynclinal rocks were thrust westward over their equivalent flat-lying shelf facies. In the Oak Hill slice he showed that Lower Cambrian sediments overlies basic volcanic rocks and that they were separated from the Sutton schists to the east by an "underthrust" which he called the Brome Thrust. (This has been renamed the Brome Slide Zone in this report.) Clark also recognized the importance of attenuated equivalents of the Oak Hill sediments (Mansville facies) within the Sutton schists and postulated that the schists were partially equivalent to the Oak Hill rocks and therefore of early Paleozoic age. Booth (1950), Shaw (1958) and Dennis (1964) accepted Clark's Oak Hill sequence in general and published detailed descriptions of the lithology and paleontology of the rocks in northwestern Vermont.

The early interpretations of the field relations on the eastern flank of the Sutton Anticline were controversial. Clark and Fairbairn (1936) thought that the Bolton lavas were extruded unconformably onto Ordovician shales in the Early Silurian; Ambrose (1942) showed that they lie conformably within the shales and hence are Ordovician. Cooke (1950) challenged this evidence and maintained that there were conformable Cambrian flows, and an unconformable Devonian lava sheet. The present work and that of Doolan et al. (1982) shows that the lavas are all conformable as Ambrose (1957) maintained in a reiteration of his evidence.

Younger rocks are downfolded into these Ordovician rocks in the Sargent Bay Syncline, and Clark (1936b), Ambrose (1942) and Boucot and Drapeau (1968) established their unconformable nature and Late Silurian age.

An encyclopedic review of the stratigraphy and paleogeography of the whole region was published by Cady (1960). Osberg (1956, 1965) made an important advance by tracing the Sweetsburg Formation (upper Oak Hill Group) eastward across a depression in the Sutton

Anticline into the Ottauquechee Formation mapped northward from northern Vermont, thereby effecting a correlation of the western and eastern sequences (Fig. 1). Osberg (1965) also recognized that Sweetsburg slates are infolded into the Sutton schists in the Valcourt Syncline and this confirmed that a major syncline lay to the west of the Sutton Anticline along the line of Clark's Brome Thrust (Fig. 1).

Cady (1969) and Rodgers (1968, 1970, 1971) reviewed the paleogeographic and structural history of the region. Rodgers postulated an extensive carbonate bank at the continent margin flanked by a volcanic island chain, whereas Cady invoked a stratigraphic convergence toward a major geanticline (Vermont-Quebec Geanticline). Cady (1969) and Osberg (1969) determined that structures were oblique to the facies trends developed on this early geanticline. In detail, the fold geometry of the Oak Hill and Sutton sequences is more complex than the simple arrangement of major structures suggests. Systems of early and late folds, some early ones being perpendicular to the general northeasterly trend, have been reported by Christman (1959), Dennis (1960), de Römer (1961), Cady et al. (1963), Rickard (1964, 1965), Osberg (1965) and Beland (1967). Cooke (1950) explained the steep folds in the Mansonville area as being due to bending about vertical axes and Cady (1969) related the development of these and east-west cross folds to the formation of a major bend in the mountain system—his New England salient (now renamed the Quebec re-entrant by Doolan et al., 1982).

The correlation of the various structures is in dispute; Eakins (1964) and Osberg (1965) interpreted the three major folds—Sutton Anticline, Richford-Valcourt Syncline and Enosburg Falls Anticlinorium—as belonging to the late phase of deformation, whereas Rickard (1964, 1965) noted that the Sutton Anticline is younger than the Richford-Valcourt Syncline and the Enosburg Falls Anticlinorium. The age of these deformations also is in dispute; workers in Quebec (Clark, 1921; Cooke, 1955; Neale et al., 1961; Rodgers, 1971; St. Julien and Hubert, 1975), impressed by the unconformity beneath the Sargent Bay Syncline, postulated that the main deformation was Taconic (Late Ordovician). Rickard (1965) argued, on the basis of K-Ar dating of muscovite associated with cleavage development, that both phases of folding were Taconic. Osberg (1965, 1969), Dennis (1968) and Cady (1969) were more impressed with the parallelism and similarity of fold styles and argued that the structures, or at least the latest phase, were Acadian (Middle Devonian).

Various dating studies in this region by Harper (1968), Cady (1969), St. Julien and Hubert (1975), Laird et al. (1984) and Sutter et al. (1985) show that ages fall into two groups and suggest that remnants of the Taconic Orogen have resisted the effects of the Acadian Orogeny. Albee (1968) and Laird et al. (1984) showed that Devonian movements retrogressed Taconic metamorphic minerals along the junction between the two groups.

Henderson (1958) postulated that much of the Cambrian to Lower Ordovician sequence is allochthonous.

Table 1. Table of formations and stratigraphic relationships.

[illegible]

Rickard (1965) argued that the Taconic Orogeny comprised three distinct phases: i) early major gravity sliding, followed by ii) folding, thrusting and slaty-cleavage development, and iii) late arching and development of crenulation cleavage. The age of this late deformation is now in doubt; it may be Acadian (Middle Devonian). Hubert (1967), St. Julien (1968) and St. Julien and Hubert (1975) demonstrated that, to the north of this area, Cambrian and Ordovician sediments were emplaced by early gravity sliding during the Taconic Orogeny to form the Stanbridge and Granby nappes (Fig. 1). This structure has been confirmed by deep seismic profiling (Laroche, 1983; St. Julien et al., 1983).

In the more recent works on this region the geological events have been placed in the context of plate tectonic theory (Bird and Dewey, 1970; St. Julien and Hubert, 1975; Poole, 1976; Osberg, 1978; Schenk, 1978; Stanley and Ratcliffe, 1985). The initial paleogeographic situation of the Quebec re-entrant is now interpreted as a triple junction with failed arm extending westward into the Precambrian basement in a zone through Montreal and Ottawa (Rankin, 1976; Doolan et al., 1982). The westernmost metabasalts (Tibbit Hill) are interpreted as early rift volcanic rocks (Kumarapeli et al., 1981; Coish et al., 1985; Pintson et al., 1985). The ultramafics and basic volcanic rocks of the eastern sequence have been interpreted as convergent-margin ophiolites (Laurent, 1975) with island arc volcanic rocks on top of oceanic basalts (Laurent and Hébert, 1978). The black shales and breccias associated with these ophiolites have been interpreted as ocean-floor melange deposits obducted onto the continental sequence to the west during the Taconic Orogeny (St. Julien and Hubert, 1975). The ultramafics are believed to mark the line of suture—the Bai-Verte-Brompton line—of the closed ocean basin (Laurent, 1975; Williams and St. Julien, 1982). The oceanic character of these ophiolites has been confirmed by detailed chemistry (Clague et al., 1981; Shaw and Wasserburg, 1984; Coish et al., 1986).

Differences in structure along the belt have been attributed to subduction-imposed compression against an irregular continental margin (Dewey and Kidd, 1974); in particular, the Quebec re-entrant strongly influences nappe development, the preservation of the ophiolite suture belt and later structures (Doolan et al., 1982; Stanley and Ratcliffe, 1985).

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STRATIGRAPHY

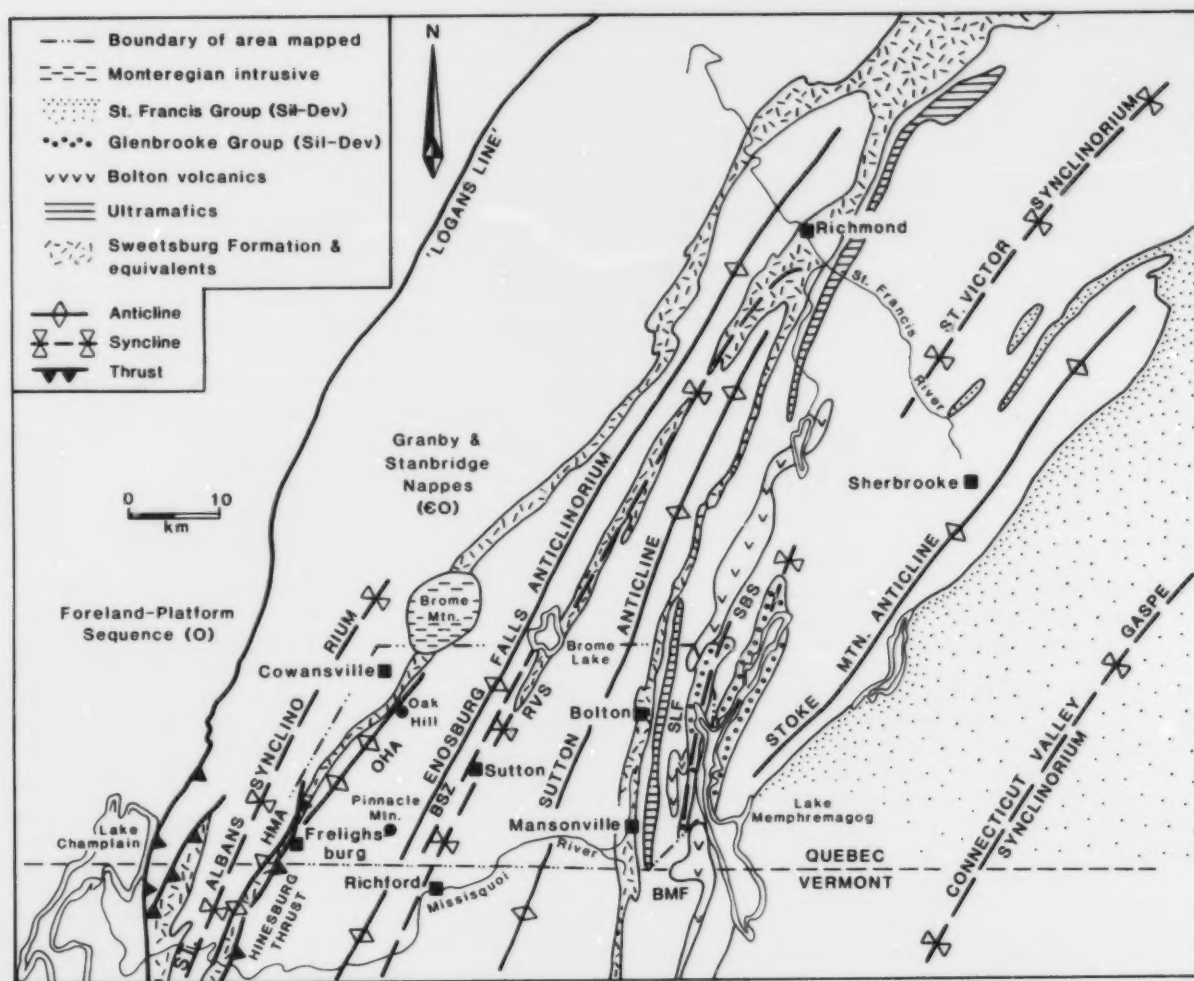
The rocks of the Cowansville-Sutton-Mansonville area can be divided naturally into three zones, the Western, Central and Eastern zones, and although Middle and Upper Cambrian strata have been traced around the northeast-plunging Enosburg Falls Anticlinorium and Sutton Anticline (Osberg, 1965), sufficient differences in stratigraphy and structure exist to warrant separate consideration of the rocks in each zone (Table 1). Correlations and a paleogeographic reconstruction are presented after discussion of the stratigraphy.

The Western Zone includes part of the Middle Ordovician Stanbridge Formation and the Hadrynian to Upper Cambrian Oak Hill Group in the Enosburg Falls Anticlinorium. The Central Zone comprises the Sutton schists, with formations equivalent to the Oak Hill Group and probably older rocks. These occupy the Richford-Valcourt Syncline and Sutton Anticline (Fig. 1). The sequence in the Eastern Zone comprises Cambrian and Ordovician sediments and volcanic rocks that outcrop to the east of the Sutton Anticline in a zone of steeply-plunging folds. Included with this sequence are the Silurian sediments in the Sargent Bay Syncline.

The stratigraphic nomenclature can only be described as chaotic, partly because of different treatments across the International Boundary and the late application of the Code of Stratigraphic Nomenclature. Recent correlations across the International Boundary have shown the equivalence of units that have long had local names and several changes have been published that are contrary to the Stratigraphic Code. The nomenclature used in this paper is set out in Table 1; the correlation problems and justification for the use of these terms are discussed in each section.

Western Zone

The Western Zone includes part of the Stanbridge Nappe and the main outcrop of the Oak Hill Group eastward to the Brome Slide Zone on the western limb of the Richford-Valcourt Syncline (Fig. 1).



OHA = Oak Hill Anticline;
 RVS = Richford-Valcourt Syncline;
 BMF = Bear Mountain reclined fold;
 BSZ = Brome Slide Zone.
 HMA = Hunter Mills Anticline;
 SLF = Sugarloaf reclined fold;
 SBS = Sargent Bay Syncline;

Figure 1. Regional geology and major structures (modified from Cady, 1960).

Oak Hill Group

Originally established as the Oak Hill Series (Clark, 1934, 1936a), the formations of the Oak Hill succession (Table 1) became accepted as a type sequence (Booth, 1950; Cady, 1960; Charbonneau and St. Julien, 1981), and they were given formal group status by Eakins (1964). Highly attenuated varieties of the Oak Hill Group lithologies in complex infolds along the eastern side of the Enosburg Falls Anticlinorium are designated the Mansville facies (Mansville phase of Clark, 1934).

Tibbit Hill Formation

The Tibbit Hill volcanics, outcropping in the core of the Enosburg Falls Anticlinorium, are the basal formation in the group. Although Dennis (1964) reduced them to member status because greywacke underlies them in Ver-

mont, others (Osberg, 1965; Charbonneau, 1980; Charbonneau and St. Julien, 1981) have continued to treat them as a separate formation.

The type locality is at Tibbit Hill, south of Brome Lake (Clark, 1936a). The volcanic rocks outcrop from Richmond in the north to North Underhill in Vermont in an anticlinal zone over 140 km long and up to 6 km wide. They are best exposed in the roadcuts 3 km north of Sutton. The formation is made up of a varied suite of basic lavas and tuffs, with local keratophyre and rhyolite flows. It is probably about 300-600 m thick. All of the basaltic types are extremely phyllitic and have been reduced by metamorphism to a chlorite-albite-epidote-calcite mineral assemblage.* Original textures and stratifi-

* Minerals are listed in order of decreasing abundance throughout this report.

cation have been almost completely destroyed by penetrative deformation and metamorphism, but rare indications of primary layering are suggested by bands of amygdules and thin epidote-rich layers. A common feature is an irregular "wire mesh" joint pattern somewhat resembling pillows; true pillow lavas occur at one locality in Vermont (Dennis, 1964). Knots and augen-like lenses of epidote and veinlets of feldspar with quartz, epidote, chlorite and in some places dolomite are common. From their mineralogy and chemistry it is clear that these rocks are largely of basaltic origin (Coish et al., 1985; Pintson et al., 1985—see section on Metamorphism and Table 2, column 7).

Metabasalt makes up the bulk of the formation. It is composed of fine grained (0.01-0.03 mm) schistose mosaics of untrinned plagioclase and chlorite with slightly larger plates of twinned feldspar. Small crystals and ragged laths of iron oxide (hematite mostly) and granules of sphene and epidote occur throughout but are usually concentrated in thin stringers parallel to the schistosity. Plagioclase (albite-oligoclase) constitutes from 25-60% of the rock, chlorite 20-50%, epidote 0-25% and sphene and iron oxide from 3-10% each. Quartz makes up to 10% and forms small lenses within the matrix. Carbonate (dolomitic) is commonly present, and accessory apatite is seen in most of the thin sections. Small irregularly rounded or streaky blebs of feldspar with epidote or quartz cores and chloritic rims were former amygdules.

Rare remnants of altered calcic plagioclase (An₆₀) suggest that the present sodic nature of the feldspar was not an original feature of the lava. A chemical analysis (Table 2, column 7) indicates that the original composition was probably alkali basalt. More detailed petrographic descriptions are given by Booth (1950) and Christman (1959).

In the upper part of the formation, especially in the northeastern part of the area, tuffaceous sediments intermediate in composition between the basic volcanic rocks and the Pinnacle greywackes are interspersed with the metabasalts. Grey slate and slaty tuff layers represent muddy sediments with admixed volcanic detritus. They are composed mainly of muscovite (sericite) and chlorite (up to 80%) with plagioclase (up to 40%), quartz (10%), iron oxide (15%) and sphene and epidote (together up to 10%). These slaty rocks are structurally important since tuffaceous fragments composed of chloritic, hematitic or talcose material, and epidote-feldspar blebs are aligned to form "willow-leaf" strain markers in the cleavage planes (see Fig. 12d). These slate layers are indicated on Map 1750A (in pocket); some of them may be infolds of the Call Mill Member of the Pinnacle Formation.

Minor thin lenses of hard grey and white finely laminated feldspathic rock occur throughout the volcanic rocks. They consist of albite-oligoclase (65%), quartz (10-15%), hematite (12%), sphene (8%), calcite (3%) and chlorite, epidote and muscovite (5%). The strong lamination is due to alternations of plagioclase-quartz layers with finer grained plagioclase mosaics containing hematite and sphene. Blebs of epidote and irregular

streaks of white feldspar or calcite distort the lamination. The rocks probably originated as local keratophyric lava flows.

The presence of rhyolites in the Tibbit Hill volcanics to the north of this area was reported by Osberg (1965) and Cady (1969). Around Mansville (see Map 1750A), lenses of light grey to white porcelaneous-weathered massive but strongly jointed rhyolite stand out prominently from the green volcanic rocks. Hematitic speckling and irregular branching meshworks pervade the rock. Although very different in outcrop appearance, this rock has a similar mineralogy to the keratophyre. Twinned and untrinned oligoclase (75%) forms a ragged slightly schistose intergrowth with quartz (15%) and this is peppered with granules of sphene and hematite. There are a few phenocrysts of subhedral plagioclase (twinned and untrinned); staining revealed no potassium feldspar. A search for zircons for U-Pb dating from several lenses of rhyolite between the latitudes of Sutton Junction and Brome failed (W.H. Poole, pers. comm., 1979). Rankin (1976) analyzed a similar rock from Waterloo to the north of this area and showed that silica is low at 69% (see Table 2, column D).

Pinnacle Formation

The original Pinnacle Formation consists mainly of well-bedded subgreywackes, but a distinct slate (the Call Mill Member) mapped by Clark (1934) as a separate formation below the greywackes is now included within the Pinnacle Formation by several workers (Dennis, 1964; Osberg, 1965; Clark and Eakins, 1968).

The type locality of the greywacke is in the synclinal inlier northeast of Pinnacle Mountain (Clark, 1936a). The greywacke outcrops as a narrow strip along the western edge of an anticlinal outcrop of the Tibbit Hill volcanics, in the synclinal outliers between North Sutton and Pinnacle Mountain—where it is best exposed—and as sheared lenses in the Mansville facies along the eastern flank of the Tibbit Hill volcanics. Its maximum thickness in the area is about 125 m; it thins in all directions, and is missing altogether in some sections, but thickens again considerably in Vermont where it also occurs below the Tibbit Hill volcanics (Dennis, 1964).

The typical rock is a medium grained greenish or grey subgreywacke. Unsorted subangular to subrounded quartz and plagioclase grains with quartz mosaic granules are set in a recrystallized matrix of muscovite, chlorite and quartz; most grains are aligned across the bedding to form a rough slaty cleavage or schistosity. In some specimens, micas and quartz of the matrix have begun to replace the clastic grains. The quartz is intensely strained and includes lines of bubbles and dust. Grain size is usually less than 0.5 mm but in a few layers is up to 3 mm or even small pebble size. Lithic fragments apart from quartz and quartz mosaic are rare. At the northern end of the Call Mill Syncline (see Fig. 5) the greywacke is pebbly but the coarse breccias or conglomerate beds that occur in Vermont (Booth, 1950; Dennis, 1964) are nowhere developed.

Table 2. Chemical analyses

	1	A	2	3	B	4	C	5	6	7	8	9	10	D
SiO ₂	59.6	58.10	64.6	67.2	66.7	78.6	77.1	48.0	42.0	46.1	47.1	50.9	50.0	68.9
TiO ₂	1.3	0.65	1.8	0.55	0.6	0.29	0.3	1.4	1.2	2.6	1.2	1.2	1.7	0.37
Al ₂ O ₃	21.8	15.40	15.6	14.6	13.5	7.0	8.7	24.4	25.5	15.9	19.3	15.0	11.9	12.0
Fe ₂ O ₃	3.1	4.02	5.8	1.1	1.6	0.5	1.5	2.0	2.2	8.5	2.3	6.5	8.7	5.4
FeO	2.6	2.45	1.8	3.4	3.5	1.7	0.7	10.0	9.3	3.4	7.4	5.7	6.4	0.91
MnO	0.02	-	.01	0.04	0.1	0.04	0.2	0.31	0.09	0.19	0.17	0.22	0.21	0.16
MgO	1.2	2.44	1.5	2.1	2.1	1.2	0.5	3.5	5.2	6.6	5.3	5.88	5.7	0.5
CaO	0.53	3.11	0.53	1.5	2.5	4.7	2.7	0.73	0.67	9.0	7.5	5.7	11.0	1.60
Na ₂ O	2.0	1.30	2.5	2.5	2.9	1.1	1.5	1.0	2.0	2.8	2.8	2.3	1.9	2.17
K ₂ O	5.3	3.24	4.5	2.6	2.0	1.6	2.8	4.4	4.0	1.4	1.6	1.3	0.12	5.77
P ₂ O ₅	0.12	0.17	0.06	0.83	0.2	0.10	0.1	0.25	0.14	0.53	0.10	0.16	0.15	0.06
H ₂ O [±]	2.2	5.00	1.2	1.9	2.4	0.6	0.9	4.9	5.6	3.5	4.0	3.6	2.1	0.87
CO ₂	0.2	2.63	0.7	0.1	1.2	2.7	3.0	nd	0.1	0.1	1.7	1.7	0.9	1.18
C	nd	0.80	nd	1.0	0.1	0.2	-	nd	1.3	nd	nd	nd	nd	nd
TOTALS	100.0	99.95	100.6	99.4	99.4	100.3	100.0	100.9	100.5	100.6	100.5	100.1	100.8	99.59
1. Lower Gilman phyllite (bulk of 5 specimens). Location—2.3 km east-southeast of East Dunham (794967) 2. Pinnacle greywacke. Location—1 km south of Mansville (876025) 3. Bonsecours graphitic phyllite. Location—1.2 km east-northeast of Abercorn (856896) 4. Sallys Pond albite gneiss (bulk of 3 specimens). Location—Highway 105A, 0.6 km north of East Richford (903874). 5. Bonsecours schist (without quartz veinlets). Location—1.8 km east-northeast of Abercorn (862897) 6. Mohawk Lake graphitic albite schist. Location—5 km east of Sutton (925978) 7. Tibbit Hill greenschist. Location—1.2 km southwest of Mansville (871026) 8. Lawrenceville greenschist. Location—3.1 km south of Brome (905040) 9. Rosenberry greenschist (bulk of 3 specimens). Location—Townline Road, 4 km south-southeast of Brome (916032) 10. Sutton Mountain albitic greenschist. Location—1 km southwest of Sutton Mountain Hamlet (889876) A. Average shale B. Average greywacke C. Average arkose A, B and C from Pettijohn (1975) D. Average of 3 rhyolites from Tibbit Hill volcanics, from Rankin (1976)														

Stratification is marked by thin magnetite-rich layers and laminae (Fig. 2), and by coarser micaceous quartzite beds and thin pelitic layers. The brown-weathering magnetite layers are most prominent to the west, especially at West Sutton where they form beds several centimetres thick which have been prospected for iron ore. The thin magnetite layers show intense contortions which contrast markedly with the gently folded but strongly cleaved quartzite in the interbeds (Fig. 2c). These black layers represent detrital heavy mineral sands which contain sphene, zircon, a few grains of labradorite and possibly rutile, as well as magnetite (see Fig. 24). Overgrown detrital tourmaline is common in the lighter coloured layers (Rickard, 1964). Such an unstable assemblage can hardly have been transported very far and yet the tiny grains are extremely well rounded.

The composition is somewhat variable; quartz varies from 30-70% (average 50%), plagioclase (albite-oligoclase) 0-35%, muscovite 5-45%, chlorite 0-35% and magnetite 0-10%. In certain layers magnetite makes up 50%, accompanied by sphene (25%) and zircon (5%). Chemical analysis indicates that the composition is close to an average greywacke (see Table 2, columns 2 and B) and to the average composition of the Canadian Shield Grenville Province (Eade and Fahrig, 1971) from which it is most likely derived. The rocks of this formation are

richer in iron and titanium, no doubt from the detrital minerals, richer in K₂O and poorer in CaO and Na₂O than an average greywacke; the latter elements probably reflect the granitic nature of the source.

The lowermost greywackes, especially in the north-east, are grey or green and white and finely laminated, similar to some of the Tibbit Hill volcanics, but the greywackes contain detrital zircon and tourmaline, whereas the volcanic rocks contain accessory apatite. The lamination is caused by alternate concentrations of chlorite and feldspar; usually it parallels bedding but in some places it is due to metamorphic segregation along cleavage planes. The greywackes become more pelitic to the east in the Mansville facies. The upper part of the formation grades sharply into the White Brook Dolomite Formation through a micaceous dolomitic quartzite which in several places contains lenses (up to 2 m by 0.75 m) of dolomite. This transitional facies is never more than a few metres thick, but it becomes more prominent to the north as the greywacke thins out.

Sedimentation features include small-scale cross-bedding which is usually extremely irregular and contorted, particularly in the magnetite layers (Fig. 2a,b,d). Lensing and small washout channels are also common (Fig. 2b,c). Reversals of cross-lamination (Fig. 2a)

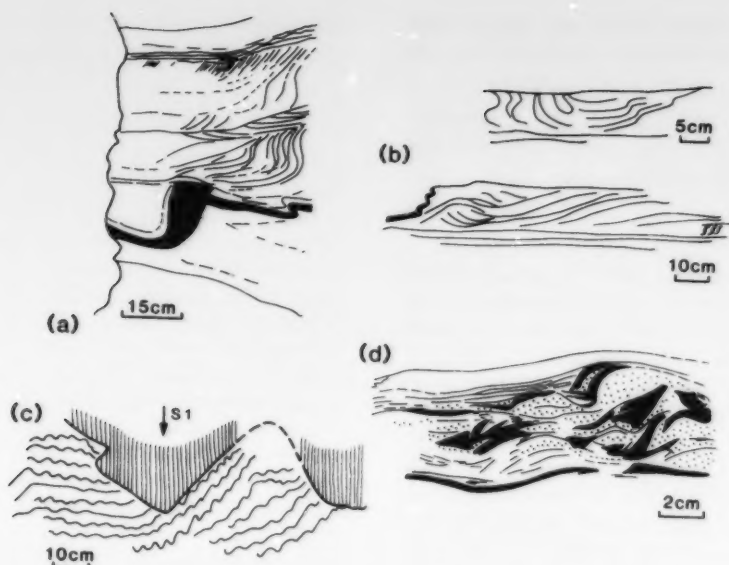


Figure 2. Sedimentary structures in Pinnacle greywacke. Black layers are detrital magnetite-rich sands. **a)** Cross-bedding. Location—2.3 km west-southwest of West Sutton (815967). **b)** Channel fill and cross-bedding. Location—0.8 km southeast of North Sutton (846012). **c)** Channel fill showing contrasting deformational behaviour between microfolded black sand layers and unlayered cleaved quartzite. Location—1 km north of West Sutton (831991). **d)** Irregular cross-bedding. Location—Perkins Road, 2.8 km west-northwest of Alva (829935).

indicating rapidly changing current directions are typical. Cross-bedding gives the "younging" direction in many places. Few graded beds or other structures typical of turbidity current deposition were found.

These rocks differ from those in Vermont; differences include the presence of magnetite-rich layers and cross-bedding, and an absence of lithic fragments and potassium feldspar in the Pinnacle Formation. In Vermont, graded beds and coarse breccias are more common (Booth, 1950; Dennis, 1964). There is no evidence to indicate reworking of older sediments in the geosyncline, and except for the lower beds the composition precludes a volcanic derivation. The mineralogy indicates a metamorphic or igneous provenance; the magnetite may well be derived from granites since magnetite-rich granite pebbles occur in the conglomerate facies in Vermont (Booth, 1950). A westward derivation is indicated by an increase in shallow water features (differential sorting, rounding and irregular cross-bedding) to the west. The Precambrian rocks of the Adirondacks are the most likely source (Booth, 1950; Cady, 1960) and such a source is consistent with their chemistry. The sedimentational characteristics have not been studied in detail but they are consistent with an origin in the proximal portion of scarp fans and aprons.

The age of the Pinnacle Formation is uncertain but its considerable stratigraphic distance below Early Cambrian fossils suggests that it may be Hadrynian, and it is so designated on the map and in Table 1.

Call Mill Member. The Call Mill strata were named as a formation by Clark (1934, 1936a) from a type locality on a tributary of Yamaska River 4 km east of West Brome. Later, Clark and Eakins (1968) reduced the Call Mill to member status. The Call Mill Member has a maximum thickness of 30 m and outcrops between the Tibbit Hill volcanics and Pinnacle greywacke throughout most of the area, but is missing in some places to the north and south. The unit has not been recognized to the south in Vermont, except by Cady (1960), or to the north; slates that do occur have been included with the Pinnacle greywacke (Booth, 1950; Osberg, 1965).

The Call Mill Member consists of dark to light grey slate in which stratification is vaguely marked by thin grey and fawn slate, fine grained quartzite or chert and sparse coarser greywacke or tuff layers. Thin quartz veinlets and zones of talcy alteration cut across the slaty cleavage possibly following or replacing the original bedding. Where it can be seen, the bedding is everywhere strongly folded and usually sheared out into lenses along the cleavage planes (see Fig. 12a). Shredded light grey and green layers are common and at Call Mill a close-spaced cleavage banding is developed (see Fig. 12h). The similarity of Call Mill slate to slate layers in the Tibbit Hill volcanics, plus the occurrence of layers and lenses of chloritic and talc-rich phyllite and disseminated hematite in Call Mill slate suggest a volcanic component. On the other hand the overall composition—muscovite (sericite) (40-90%), chlorite (20%), plagioclase (10%), quartz (20%) and hematite and sphene (15%)—and more especially the detrital

accessory minerals tourmaline, zircon and sphene, indicate a derivation from a generally nonvolcanic source.

The deposition of these fine muds with possibly some volcanic admixture marked the cessation of volcanic activity and the commencement of rapid clastic sedimentation of the Pinnacle Formation.

White Brook Formation

The White Brook dolomite was treated as a formation by Clark (1934, 1936a) and retained its formational status in some later works (Booth, 1950; Eakins, 1964; Clark and Eakins, 1968). It has been reassigned as a member of the Underhill Formation (Doll et al., 1961; Dennis, 1964) and of the Pinnacle Formation (Osberg, 1965, 1969). The value of this thin dolomite marker for mapping structural complexities justifies its formational status, and is treated as such in this paper.

The dolomite is best exposed near the type locality at West Sutton (Clark, 1936a). It varies from 30 m thick to about 3 m or less in the Mansville facies; it has been traced around the closure of the Enosburg Falls Anticlinorium (Osberg, 1965) and occurs within most of the synclinal outliers of Pinnacle greywacke. It is commonly missing in Vermont but it has been traced intermittently south of the International Boundary for 25 km to Fairfield (Booth, 1950; Doll et al., 1961; Dennis, 1964). The lower boundary with the Pinnacle greywacke is transitional and the upper contact is sharp in places but elsewhere transitional into a hematite rock that marks the base of the Gilman Formation.

The dolomite is interlaced with irregular quartz veinlets and joint fillings that give it a superior resistance to erosion. The typically brown-weathered surface is stippled with grains and clumps of quartz crystals and cut by platy stringers of chlorite. The fresh surface is mottled in white, grey, brown and pink and speckled with quartz and chlorite. Quartz-rich layers and thin fine grained white calcite streaks and veins may, in places, indicate primary bedding. Bedding is rare but chlorite and muscovite commonly lie in discrete parting planes giving a rough schistosity parallel to the regional cleavage. The rock is a fine to medium grained quartzitic dolomite marble. Quartz locally makes up to 50% of the rock in the lower portion of the formation where it occurs in large (1-2 mm) grains or pebbles of quartz mosaic. The single grains are intensely strained and contain thin zones of included bubbles; the mosaic consists of clear relatively unstrained quartz. In many cases the single grains are rounded clastic grains, although their margins are embayed by the matrix. Toward the east the clastic texture is modified by metamorphism. A small amount of sodic plagioclase is also present. The character of the clastic grains suggests derivation from the same source as the Pinnacle greywacke. The matrix consists of fine grained (0.05-0.5 mm) equigranular dolomite, quartz and calcite with muscovite and chlorite in some thin sections. Accessory minerals include iron oxide, zircon and tourmaline.

Gilman Formation

The Gilman Formation comprises phyllite, siltstone and quartzite and includes in its basal part the Alderbrook ferricrete and West Sutton slate members. Clark (1934, 1936a) recognized that the upper quartzitic part of the formation was a correlative of the Cheshire quartzite in Vermont (Keith, 1923) but chose a new name because the lithology was sufficiently different. Although Booth (1950) used the name Gilman in Vermont, the Cheshire has now been extended north to the International Boundary (Doll et al., 1961; Dennis, 1964) and into southern Quebec (Charbonneau, 1980). Clark (1934, 1936a) included a thick unit of siltstone and phyllite in the lower part of his Gilman Formation, and recognized an underlying West Sutton slate formation. Booth (1950) included most of this lower Gilman with the West Sutton, and Dennis (1964) placed both together in the Fairfield Pond Member of the Underhill Formation. Osberg (1965, 1969) retained the name Gilman for the upper part of the sequence in the area to the north but correlated the lower unit with part of the Sutton schists which he renamed Bonsecours Formation. Charbonneau (1980) unnecessarily introduced a new name, Frelighsburg Formation, for the lower part of the Gilman. Marquis (1985, 1986) also used the term Gilman for all of the phyllites between the Pinnacle and Sweetsburg formations in the Richmond area. There are objections to direct correlation with the Sutton schists since they probably contain additional units, so that the terms Underhill and Bonsecours are not used here in the Vermont Zone. The simplest compromise used in this report and map is to recognize upper and lower parts of the Gilman Formation following Clark and Eakins (1968) and include the West Sutton as a member of the Gilman following Booth (1950).

Alderbrook Member. Clark (1936a) and Booth (1950) recorded the occurrence of hematite at the base of the West Sutton slate. The hematite rock reaches a maximum thickness of only 1.2 m but it is a remarkably persistent marker bed useful in mapping complex folds. The hematitic rock is accordingly here named the Alderbrook Member. The type locality is to the east of Alderbrook Road, 1.2 km south of North Sutton.

The rock is a dense maroon to greenish-grey slate very rich in hematite (up to 60%). A dense chlorite-rich rock containing dolomite lenses commonly interlayers or veins the hematite rock. The slate also contains as much as 30% each of angular quartz and plagioclase, and some dolomite and magnetite. The junction with the underlying White Brook dolomite is in some places sharp and elsewhere transitional where the hematized dolomite has a brecciated texture and the margins of the fragments have been replaced by hematite. Hematite- and chlorite-rich ribs (about 7 cm thick) penetrate the dolomite. Rounded reddish quartz grains and quartz-mosaic pebbles, similar to those in the dolomite, occur in the hematite- and chlorite-rich rock in several places; these appear to be remnants from the replacement of the dolomitic matrix.

As suggested by T.H. Clark (pers. comm. 1957), this hematite layer is interpreted to be a paleoferricrete lying on the White Brook dolomite. That this layer marks a disconformity is indicated by the presence of the hematite between the Pinnacle greywacke and West Sutton slate where the White Brook dolomite is missing (Booth, 1950). The overlying West Sutton slate has disseminated hematite in its basal part; it marks a marine transgression after the erosion.

West Sutton Member. The West Sutton Member (Clark and Eakins, 1968) outcrops in a 1.6 km-wide zone west of West Sutton, its type locality, where it has a maximum thickness of 80 m. To the north and south it thins abruptly to less than 3 m and in the synclinal outliers it is missing or very thin. Generally, it has not been recognized outside the Sutton area. The lithology is similar to that of the Call Mill Member—medium to dark grey slate with some thin white or light grey layers and thin irregular talcose laminae—except that coarse pebbly greywacke beds (2.5–5 cm thick) occur at some localities, and disseminated hematite produces a characteristic reddish streak.

Muscovite-sericite constitutes 80–90% of the rock with minor quantities of quartz, albite-oligoclase, iron oxide (hematite and magnetite), chlorite and accessory calcite, tourmaline and zircon. The chlorite and magnetite are porphyroblastic in some layers. The member passes transitionally upward into Gilman phyllite.

Gilman Formation, lower part. Gilman phyllite and siltstone outcrop poorly over a 6 km-wide belt; the unit has been traced for many kilometres to the north and south. Clark's (1936a) type locality at Oak Hill is in the upper part of the formation; a suggested type locality for the lower part is the area between West Brome village and Gilman. Structural considerations (enveloping surfaces to folds) indicate that the outcrop is essentially a horizontal sheet in spite of the local steep dips (see cross section on Map 1750A). The maximum thickness is probably about 400 m.

The phyllites vary in colour from shades of light grey and greenish grey to silver; a white-weathered crust and rusty stains on cleavage planes are characteristic. Pyrite cubes as large as 5 mm are ubiquitous, and small magnetite grains occur in some localities, especially in the synclinal inliers to the east. Bedding is represented by thin siltstone layers which have been fractured and displaced along cleavage planes; thin layers are preserved near the base and are intricately folded. The silt-sized quartz grains are set in a micaceous matrix and the clastic texture is still visible in some of the siltstone lenses in spite of the strongly developed micaceous cleavage. Toward the east, where the metamorphic grade is higher, thin "segregation" quartz veinlets lie in the bedding and cleavage planes. The average composition is quartz (45%), muscovite-sericite (45%) and chlorite (10%); small percentages of sodic plagioclase, magnetite and pyrite are present, and tourmaline, zircon, sphene, apatite and

hematite are common accessory minerals. Chemical analysis indicates that the phyllite is close to the composition of an average shale (Table 2, columns 1 and A) or a plagioclase-rich greywacke (Crook, 1974). It is richer in Al_2O_3 than shale, and richer in K_2O and poorer in CaO than both shale and plagioclase greywacke. Some calcium and silica has probably been leached out of the system judging from the composition of the veinlets.

Toward the west the phyllites become dark grey and slightly coarser and pass transitionally up into fawn and grey silty quartzites. A dark grey phyllite occurs along the eastern side of Oak Hill.

Gilman Formation, upper part. The lower Gilman phyllites grade upward into silty quartzite that forms a low but prominent ridge extending from the Brome Mountains (north of the area) southward through Dunham and Frelighsburg into Vermont. Where exposure is poor, the transitional boundary between the lower and upper parts of the Gilman Formation is difficult to locate and the mapped boundary depends on the topographic expression of the more resistant siltstones and quartzite of the upper Gilman. Although the upper Gilman does resemble the lower part of the correlative Cheshire Formation in Vermont, it lacks the thick beds of pure white quartzite of the type section of the Cheshire. Cady (1945) has suggested that the pure quartzite offlaps the lower impure quartzite toward the south. Osberg (1965, 1969) has traced the Gilman quartzite northward as far as the St. Francis River and has mapped a thin lens 1.6 km north-east of Richmond on the eastern side of the Enosburg Falls Anticlinorium. It has not been recognized in this position anywhere to the south.

The type locality is on Oak Hill (Clark, 1936a) where the quartzite is about 300 m thick; it thins northward and pinches out at the St. Francis River. Good exposures occur in roadcuts east of Sweetsburg. The typical upper Gilman, a white-weathered fawn, grey and white fine grained impure quartzite, usually has a coarse shredded texture, with disrupted lenses of quartzite being set in a cleaved grey or black phyllite. Beds are generally reduced to an alignment of lenses but some planar interbeds of quartzite, dolomitic quartzite and grey slate, a few centimetres thick and rarely a few metres thick, can be seen near the top of the formation. Beds generally dip steeply west. Gash veins of quartz and dolomite cut some of the thicker more massive units, and wafer-thin quartz veinlets cut across and parallel the cleavage in the western upper part of the formation.

The quartzites consist of quartz (50–60%), muscovite-sericite (20–40%), feldspar (10%) and dolomite (as much as 10% in some samples), with accessory graphite, pyrite, tourmaline and zircon. Composition and grain size (0.15–0.2 mm) suggest that the original sediment was a silty sandstone. Plagioclase and microcline are usually fresh or only slightly altered; several grains show overgrowths of new feldspar. Plagioclase with deformed twin lamellae or zoning indicate a metamorphic-igneous rock provenance.

Fossils have been found in the upper Gilman at Scottsmore (Clark, 1936a) and in the Cheshire Formation (Cady, 1945; Shaw, 1954) that establish the age as Early Cambrian. These fossils are the oldest found in the Oak Hill Group.

Dunham Formation

The Dunham Formation consists of dolomite and two thin members, the Oak Hill slate and Scottsmore quartzite, which were all named by Clark (1934, 1936a) from type localities on Oak Hill. The Dunham dolomite has been traced as far southward as the western limb of the St. Albans Synclinorium (Fig. 1; Shaw, 1958) where the Oak Hill and Scottsmore correlate with the Parker and Rugg Brook formations respectively. On the western limb of the Enosburg Falls Anticlinorium, however, Oak Hill and Scottsmore are recognized only with difficulty and were grouped with the Dunham into the Bridgeman Hill Formation (Dennis, 1960, 1964; Doll et al., 1961). To the north these formations thin considerably, and Cooke (1954) combined them into a "Dunmore" formation, whereas Osberg (1965), Charbonneau (1980) and Marquis (1985), finding them too thin to map, included them with the Sweetsburg Formation. This work follows Clark and Eakins (1968) by incorporating the Oak Hill and Scottsmore as upper members of the Dunham Formation.

Combining these three units is a logical step since the Oak Hill slate is not mappable except at the type locality and the Scottsmore quartzite is thin and not everywhere lithologically distinctive. Moreover, the sequence has a lithological continuity; thin dolomite beds occur in the upper part of the Gilman quartzite, thus producing a gradation up into the Scottsmore dolomitic quartzite; black slates, precursors of the overlying Sweetsburg, also occur in the uppermost Gilman (exposed in a roadcut 0.8 km south of Sweetsburg) and in the Oak Hill slate.

The Dunham Formation outcrops in a narrow belt from Sweetsburg to Frelighsburg and in an offset belt from west of Dunham to the International Boundary. Most dips are steep in the overturned limb of the Oak Hill and Hunter Mills anticlines, and both belts bifurcate and rejoin due to a variable plunge on minor folds. Eakins (1964) joined the two belts west of Dunham but they are shown here separated by a fault; the critical outcrops are shown on a detailed map (see Fig. 10). Synclinal outliers occur at Scottsmore and southwest of Frelighsburg. The thickness varies from 50 m at Oak Hill to about 10 m west of Dunham.

The dolomite is typically massive, dark bluish grey to black and fine grained. Abundant quartz sand grains stand out on the brown-weathered surface. Irregular quartz veins and ribs are prominent, as in the White Brook dolomite. Light grey to fawn, white and yellowish varieties of dolomite, and quartzite and shaly beds occur locally especially southwest of Frelighsburg. Bedding is uncommon but is indicated in places by shaly or sandy layers (see Fig. 8f). Axial-plane cleavage affects the shaly layers and is accentuated by wafer-thin quartz veinlets. As seen under the microscope the dolomite is com-

posed of equigranular to elongate mosaics, with up to 30% rounded quartz grains. The dark colour of the dolomite is due to scattered inclusions and rims of graphite. Coarser patches of dolomite are associated with stringy veinlets of microcline. Some varieties contain mainly calcite. Accessory minerals include plagioclase, tourmaline, zircon, pyrite and in some places galena. The shaly varieties contain considerable (10-20%) sericite and chlorite, usually concentrated into spaced cleavage layers.

Clark (1936a) assigned the dolomite an Early Cambrian age on the basis of *Mesonacid* trilobite fragments found 5.8 km southwest of Frelighsburg.

Oak Hill Member. Clark (1936a) described the Oak Hill slate as a tough, dark brownish-grey or black siltstone. Stratification is shown by limonite flecks, and minor discordances are common. It is only a few metres thick and could not be traced away from the type locality.

Scottsmore Member. The Scottsmore Member is typically a laminated and locally cross-bedded dolomitic quartzite. Massive siliceous and conglomeratic varieties occur in a few lenses. The maximum thickness is 8 m (Clark, 1936a) and the member is more conspicuously exposed along the gentler dipping fold hinges than the narrow belts of steep dip. Charbonneau (1980) reports repetitions of the Scottsmore facies in the Sweetsburg slate north of St. Armand Centre but these were mapped as structural infolds during the present work.

Specimens of fawn-and-white mottled quartzite examined microscopically contain 70% angular quartz grains, 15-20% dolomite and 5-10% feldspar. Plagioclase occurs as prisms and microcline as rounded grains; some have overgrowths. In addition to quartz grains, the conglomerates contain fragments of reworked dolomite and limestone. Osberg (1965) recorded Early Cambrian fossils in the Scottsmore to the northeast of this area.

Sweetsburg Formation

First described by Clark (1936a) from a type locality on Oak Hill, the Sweetsburg Formation outcrops in a 1.5 km-wide strip in westward-overturned synclines to the west of the Oak Hill and Hunter Mills anticlines. The formation has been traced south of the International Boundary to St. Albans and has been correlated with the St. Albans, Skeels Corners and Hungerford slates in the western limb of the St. Albans Synclinorium (Fig. 1; Shaw, 1958; Cady, 1960); northward the formation closes around the Enosburg Falls Anticlinorium, Sutton Anticline and intervening Richford-Valcourt Syncline in the vicinity of Richmond and thence has been traced southward as the equivalent Ottawaquchee Formation of the eastern sequence (Osberg, 1956, 1965; Cady, 1960). The Sweetsburg is much thicker than Clark's (1934, 1936a) estimated 100 m and is more likely 500-600 m.

The lower Sweetsburg has a characteristically uniform lithology of black pyritic slate with interbedded laminae of white to grey siltstone (quartzite). Rusty or yellow

alteration of the cleavage surfaces is common and finely developed slaty cleavage is in places almost obliterated by a later crenulation cleavage. The rock is very fine grained and consists of about 80% muscovite-sericite, 5% each of chlorite and graphite dust and 5-10% quartz. Pyrite is ubiquitous and in some specimens accessory detrital tourmaline is overgrown to form tiny prisms (Rickard, 1964). The silty laminae consist mainly of quartz with some sericite and a little plagioclase.

The widespread development of this rock type indicates a marked change from previous sedimentary conditions. The formation is assigned a Middle Cambrian age on the basis of its correlation with fossiliferous formations in the western limb of the St. Albans Synclinorium (Shaw, 1958; Cady, 1960).

Vail Member. Distinguished by Clark (1936a) as an upper member of the Sweetsburg Formation, the type locality of the Vail Member is at Vail Cemetery on the road west of Oak Hill. The unit has been recognized elsewhere only in the western limb of the St. Albans Synclinorium as the Hungerford slate (Dennis, 1964). It differs from the underlying slate in its more conspicuous bedding, and consists of thinly interbedded bluish-black, grey and yellowish-grey slates with light grey siltstones and sandy beds (some dolomitic) up to 15 cm thick (rarely 45 cm thick). Weathering accentuates the bedding and minor structures well (see Fig. 8a,b,c).

This member may be as thick as 300 m in places. Its lower contact is probably transitional and its distinction from the underlying Sweetsburg slates is not everywhere certain; west of Pell Pond, for example, Vail-type slates seem to lie directly upon the Dunham dolomite. Charbonneau (1980) recognized an upper "ardoise rubanee" unit to the Sweetsburg but he placed part of Clark's (1936a) Vail in the Stanbridge Formation.

Stanbridge Formation

Eakins (1964) and Clark and Eakins (1968) grouped the widespread but poorly exposed rocks west of Cowansville and east of "Logan's Line" as the Stanbridge Complex. The rocks form an allochthonous unit referred to by Clark (1934) as the Philipsburg slice and by St. Julien and Hubert (1975) as the Stanbridge nappe.

Only the eastern part of this unit was examined, where it occurs in poor low exposures west of the Sweetsburg strata. In view of the supposed allochthonous nature of this zone the lower junction is of great interest; it is unfortunately nowhere exposed in this area, although Stanbridge and Sweetsburg slates are only a few metres apart in the road section west of Meigs Corners. The lithology is similar to the Vail and can only be distinguished by the presence of thin limestone beds in the Stanbridge. The best exposure is in Rioux Quarry, Cowansville; there the strata are black to light grey phyllitic slate with thin (2-15 cm) interbeds of yellowish-grey-weathering blue limy shale and grey limestone beds (up to 0.6 m thick) and thinner siltstone laminae. Pyrite cubes and nodular radi-

ating aggregates are common. The shales and slates are cut by wafer-thin quartz veinlets, and the limestones by calcite-filled extension gashes. The age of this formation is established as early Middle Ordovician from graptolites in the Cowansville quarry (Riva, 1966; Williams, 1966).

Clark (1934) showed the junction between the Stanbridge Formation and the Oak Hill Group as a thrust but there is little difference in structural style across the junction (cf. Fig. 8 and 9). Mapping (Eakins, 1964; this report) suggests that the contact may be a simple disconformity, and it is noteworthy that these Stanbridge rocks are very similar to the Melbourne limestones (Cooke, 1954; Cady, 1960) near Richmond that Osberg (1965) included as an upper member of the Sweetsburg Formation and Marquis (1985, 1986) placed transitionally above the Sweetsburg as the uppermost member of the Oak Hill Group. Claude Hubert (pers. comm., 1976) suggested that the tectonic break may be farther west. On the other hand, Charbonneau (1980) suggests that the Cowansville rocks are high in the Stanbridge sequence which if confirmed, would support the existence of a major break along this boundary.

Central Zone (Sutton schists)

The Central Zone of schists can be subdivided into four major units: Sallys Pond, Mohawk Lake, Bolton Glen and Bonsecours formations (see Table 1). Stratigraphic correlation is difficult in the Sutton schist belt; the problems were discussed by Cady (1960, 1969). Clark (1934) and Cooke (1954) recognized Oak Hill equivalents in the Sutton schists and postulated that younger beds may be present. In northern Vermont the schists have been placed in the Camels Hump Group comprising Underhill, Jay Peak, Hazens Notch and Pinney Hollow formations (Cady, 1960), but the stratigraphic relationships among them are unclear (Osberg, 1969), and these units may be in separate thrust slices (Stanley and Ratcliffe, 1985). Cady (1960) and Doll et al. (1961) showed the general equivalence of the schists to the Oak Hill Group and even included some of its formations (Pinnacle and White Brook) with the schists in the Camels Hump Group. In southern Quebec, Osberg (1965) named the upper part of the schists the Bonsecours Formation, and Cady (1960), Doll et al. (1961), Dennis (1964) and Osberg (1965) demonstrated that the Underhill and Bonsecours formations were partly equivalent to the lithologically similar lower Gilman phyllites and, because of facies changes, are probably equivalent to the whole Gilman Formation in the Sutton Anticline (Osberg, 1965, 1969). Thus the Bonsecours schists are probably of Early Cambrian age. Other thin units of the Oak Hill Group occur in tectonic inliers within the schists along the western limb of the Richford-Valcourt Syncline (see map). These constitute the Mansville facies of the Oak Hill Group (Clark, 1934).

The Sweetsburg Formation overlies the Bonsecours schist in the core of the Richford-Valcourt Syncline in a continuous belt north of Brome Lake (Osberg, 1965). It continues southward through the report area along the synclinal hinge zone as lenticular outliers. Sweetsburg

slate is also associated with the Mansville facies and the structural interpretation of these rocks is critical to the correlation of the Sutton schists with the Oak Hill Group. Oak Hill rocks are not present on the eastern limb of the Richford-Valcourt Syncline—either there is a facies change, or they have been structurally removed. In any case, since the early folds in the schists are generally not large enough to affect the outcrop patterns, the sequence should become older eastward and downward from the Sweetsburg Formation in the core of the Richford-Valcourt Syncline. Therefore the central part of the Sutton schists should be equivalent to or older than the lower Oak Hill Group and are probably Hadrynian in age (cross section).

Three units below the Bonsecours schist were recognized by Osberg on his unpublished 1958 manuscript map, and in this study they were found to continue southward to the International Boundary. They are not here formally named since their status and extent are not yet clear and their contacts may be tectonic breaks. The quartzofeldspathic schists and greenschists of the Bolton Glen formation are probably equivalent to the Pinnacle Formation (Table 1). Below these are graphitic schists and greenschists of the Mohawk Lake formation and albite schists and gneisses of the Sallys Pond formation. Correlations to the south depend on the interpretation of facies changes and the linking of disconnected greenschist outcrops. The Bolton Glen formation appears to be partly equivalent to the Jay Peak Formation, and the Mohawk Lake and Sallys Pond formations probably equate with the Hazens Notch Formation (Doll et al., 1961).

Petrographic details of these schists are mainly described in the section on Metamorphism. On the map, pelitic and graphitic rock types are distinguished from semipelite and quartzofeldspathic ones for each formation.

Sallys Pond formation

The central core of the Sutton Mountains consists of massive to layered feldspathic schist and gneiss, quartzite and semipelite with minor layers of black graphitic schist and greenschist referred to informally as the Sallys Pond formation. The unit extends from Bolton Pass in the vicinity of Sallys Pond, the type locality, south to the International Boundary; its extension farther south is unknown because it passes into the graphitic Hazens Notch Formation (Doll et al., 1961). This is interpreted to be the oldest unit in the area and although its base is not exposed it must be at least 1000 m thick.

The distinction of this unit depends to some extent on its metamorphic state. The abundant feldspar porphyroblasts are usually black from included graphite dust or have black cores, indicating that the original sediments contained a significant amount of carbonaceous matter as does the overlying Mohawk Lake formation. It was also possible to map some separate lenses and thin layers of dark grey to black graphitic albite schist (see Map 1750A). Nevertheless, the preponderance of quartzite and

semipelite in layers ranging from a few centimetres to a few metres thick serve to distinguish the unit from the overlying more pelitic rocks. There are only a few lenses of green chlorite-albite-(amphibole-biotite) schist in this formation and similar metavolcanics are more abundant in the overlying Mohawk Lake schist.

The lithological layering in these schists and gneisses probably represents original bedding, and a bedding schistosity is enhanced by abundant thin quartz segregations. The rocks contain variable proportions of feldspar (albite-oligoclase) in porphyroblasts up to 3 cm long, quartz, muscovite, chlorite, graphite and apatite with minor but in places porphyroblastic calcite, magnetite, tourmaline, biotite, chloritoid and garnet. Garnet and chloritoid were observed only at Glen Sutton and are restricted to thin layers. Pyrite is common especially in the quartzitic layers. A bulk sample of three specimens of medium to coarse grain size shows that the chemistry is close to that of an average arkose (Table 2, cf. columns 4 and C) except that CaO is higher and K₂O lower in the sample; possibly these were plagioclase arkoses.

Mohawk Lake formation

Some 700 m of orange and grey, rusty-weathered black pelitic and semipelitic schist occurs with gradational contact above the Sallys Pond formation on both flanks of the Sutton Anticline. The type locality is the small lake 4.5 km southeast of Sutton. The graphitic feldspathic schist contains variable proportions of quartz, albite, chlorite, muscovite and graphite interlayered (a few centimetres to a few metres in thickness) to form an assemblage of pelite, semipelite and impure grey quartzite. Pyrite is ubiquitous and occurs in cubes up to 1 cm across. The semipelite is rich in black albite porphyroblasts. Less graphitic semipelitic schist (distinguished on the map) becomes more common toward the south. Prominent greenschist lenses occur within the unit on the western side. The several layers in the south near the hamlet of Sutton Mountain consist of layered (1-10 cm) coarse and fine, chlorite- and albite-rich schist. A few epidote and feldspar lenses and quartz "knots" distort the schistosity, and one 10 cm-thick layer of chlorite-flecked marble was observed. Originally these greenschists were basalts (Table 2, column 10). The main layer here can be traced discontinuously southward into the Peaked Mountain "greenstone" in Vermont (Doll et al., 1961; Dennis, 1964). The Mount Echo greenschist body, 9 km east-northeast of Sutton, is probably at the same stratigraphic level; it occurs as a lenticular core of an early fold, and is unusual in that it contains massive to finely schistose, fine grained, grey and cream albite-chlorite rock in addition to the normal chloritic and albitic layered variety. These more massive rocks may represent a less deformed core of the original volcanic rocks.

Bolton Glen formation

In the north of the area the Mohawk Lake formation passes up sharply into about 300 m of quartzofeldspathic

schist followed by 150 m of greenschist. From a type locality about 1 km southwest of Bolton Glen these two units outcrop along both flanks of the Sutton Anticline. The quartzofeldspathic schists include three typical rock types: light green quartz-chlorite-muscovite schist; grey-green albite-chlorite-quartz-muscovite schist; and grey quartzite beds up to 1 or 2 m thick. Conspicuous but minor layers of graphitic schist and lenses of limestone are indicated on the map. All of these rocks are inter-laminated with thin quartz veinlets parallel to a strong bedding schistosity.

The greenschist is an important marker horizon informally referred to as the Rosenberry member, from a type locality north of Rosenberry Road and 3 km northeast of Sutton Junction. The Rosenberry member is a chlorite-albite-epidote-calcite-colourless amphibole-muscovite-magnetite schist with chlorite- and albite-rich layers and a few interbeds of metasediment. It is of basaltic composition (Table 2, column 9). Some pods of serpentinized peridotite also occur within the unit and may mark the presence of a thrust (discussed later).

On the eastern limb of the Sutton Anticline, the Bolton Glen formation and the Rosenberry member can be traced discontinuously to the International Boundary and it is well exposed on Highway 105A 1 km west of Highwater. A separate inlier of this unit together with peridotite and marble lenses occur 2 km southeast of Knowlton. On the western limb, the formation can be traced to about 2 km south of Sutton where it is reduced to 2 m of platy quartzite and 1.5 m of greenschist; farther south a few outcrops of less than 1 m thick of bluish-grey streaky quartzite continue on trend toward the International Boundary into a lens of greenschist east of Richford (Dennis, 1964). Along the southern section, the contact is gradational as the Bolton Glen formation has pinched out, and graphitic schist characteristic of the Mohawk Lake formation passes with little change westward into the Bonsecours Formation.

Bonsecours Formation

This name was given by de Römer (1961) and Osberg (1965) to phyllitic schist in the hinge region of the Sutton Anticline (their St. Etienne Anticline) 30 km north of the study area. Osberg (1965) mapped a thin but important greenschist marker (Lawrenceville member) for 70 km around the north-plunging anticline. He also traced the Bonsecours schist westward around the nose of the Enosburg Falls Anticlinorium into the lower Gilman phyllites, and thus established that this pelitic unit expanded eastward from 175 m to as much as 660 m thick.

The Bonsecours Formation is poorly exposed in the Richford-Valcourt Syncline through the Sutton valley, and on the eastern flank of the Sutton Anticline west of the North Missisquoi valley. The base of the formation was not defined by Osberg (1965); in the east and north, where the formation crosses the Sutton Anticline, the base is taken as the top of the Rosenberry greenschist or the quartzofeldspathic schist of the Bolton Glen formation.

On the western side of the Richford-Valcourt Syncline there is only a tectonic boundary.

The formation consists of greenish-grey to silver phyllite with minor grey and black slate, white quartzitic phyllite and quartzite in laminae or beds up to several centimetres thick. Wafer-thin quartz veinlets are common and are usually parallel to bedding. The phyllites consist of variable proportions of muscovite (20-70%), quartz (30-80%), chlorite (up to 10%) and albite (up to 10%) with lesser amounts of magnetite, ilmenite and biotite and accessory tourmaline, sphene and zircon. Biotite is prominent only in grey phyllite layers south of Sutton. A graphitic phyllite from 2 km northeast of Abercorn is chemically close to average greywacke except for less CaO and more carbon (Table 2, cf. columns 3 and B). The lithology is similar to that of the lower Gilman Formation except that in the Bonsecours Formation, magnetite occurs as small octahedra and albite has developed as pinhead-size porphyroblasts.

Thin layers of greenschist consist of chlorite-albite-epidote-calcite schist. They are commonly segregated into chlorite-rich and albite-rich layers a few centimetres thick and have associated minor dolomite or calcite marble layers. Like the Tibbit Hill volcanics they probably originated as alkali basalt (Table 2, column 8). The small lenses near Brome close to the Sweetsburg outcrop probably represent the continuation of the Lawrenceville member (Osberg, 1965). Those in the south, near Abercorn, line up with the lenticular Bakersfield greenstone in Vermont (Dennis, 1964) and may itself be stratigraphically equivalent to the Lawrenceville.

On the eastern limb of the Sutton Anticline, the Bonsecours Formation is recognized as generally nongraphitic phyllite sandwiched between the Bolton Glen and Sweetsburg formations. Several greenschist layers or lenses up to 10 m thick have been mapped within the Bonsecours Formation; some may be repeated on folds. The uppermost layer may represent the Lawrenceville member and probably correlates with the Belvidere Mountain amphibolite to the south in Vermont (Cady et al., 1963). Several thin black carbonaceous phyllite layers also exist and some have thin interbeds of dolomite a few centimetres thick.

Mansville facies of the Oak Hill Group

The extraordinary occurrences of attenuated formations of the Oak Hill Group within the Bonsecours schists along the western limb of the Richford-Valcourt Syncline warrants special mention. The Oak Hill formations thin markedly eastward across the crest of the Enosburg Falls Anticlinorium in successive synclinal outliers; the thinning is irregular, and in places the Pinnacle greywacke and White Brook dolomite are missing so that the lower Gilman phyllites are in tectonic contact with the Tibbit Hill Formation on the eastern limbs of some synclines (see cross section).

The abrupt junction of Tibbit Hill volcanics and Bonsecours schist (equivalent to the Gilman Formation) on

the western side of the Sutton valley marks a major structural discordance—the Brome Slide Zone. The intervening units of the Oak Hill Group occur in lenses along the contact and in tight infolds within the Bonsecours schist. About 1600 m of the section on the western limb of the Enosburg Falls Anticlinorium is reduced to less than 70 m in the lenses on the eastern limb of the anticlinorium. Clark (1934) referred to this reduced sequence as the “Mansville phase” after a type locality at Mansville, 5 km north of Sutton. It is useful to retain this name as the Mansville facies to refer to all of the outcrops of the thinned Oak Hill Group strata in the Richford-Valcourt Syncline. Wherever possible the beds were mapped as individual named units, but where it was impossible to separate two or more lithological units they are shown on the map as undivided Oak Hill Group.

The Pinnacle greywacke is commonly represented by its calcareous upper portion or by laminated and schistose varieties that are difficult to distinguish from the Gilman phyllite. The presence in sequence of such distinctive lithologies as Tibbit Hill greenschist, Call Mill slate, Pinnacle greywacke, White Brook dolomite and Alderbrook hematitic siltstone, even where each unit is

only a few metres thick or, in the case of the Alderbrook, only a few centimetres thick, enables the sequence and structure to be determined wherever the exposure is adequate. A large-scale map (Fig. 3) shows details of the main outcrop from Mansville to Sutton; structural details are considered later. The zone of the main development of the Mansville facies continues southward as isolated lenses. Formations as young as the Sweetsburg occur in this zone (Clark, 1934; unpublished map, 1931); the outcrops southwest of Brome include grey and fawn dolomite and black laminated slate. Similar black phyllites with carbonate lenses outcrop to the south especially around Abercorn (see map 1750A). These are probably attenuated versions of the Sweetsburg and Dunham formations, and this interpretation has important stratigraphic implications because Cady (1960) and Dennis (1964) have designated similar dolomites in black slate along the eastern margin of the Tibbit Hill volcanics in Vermont as White Brook dolomite, which is lower in the sequence.

The interpretation of the Mansville facies is difficult. There are undoubtedly sedimentary changes with a general increase in finer grained sediments to the east,

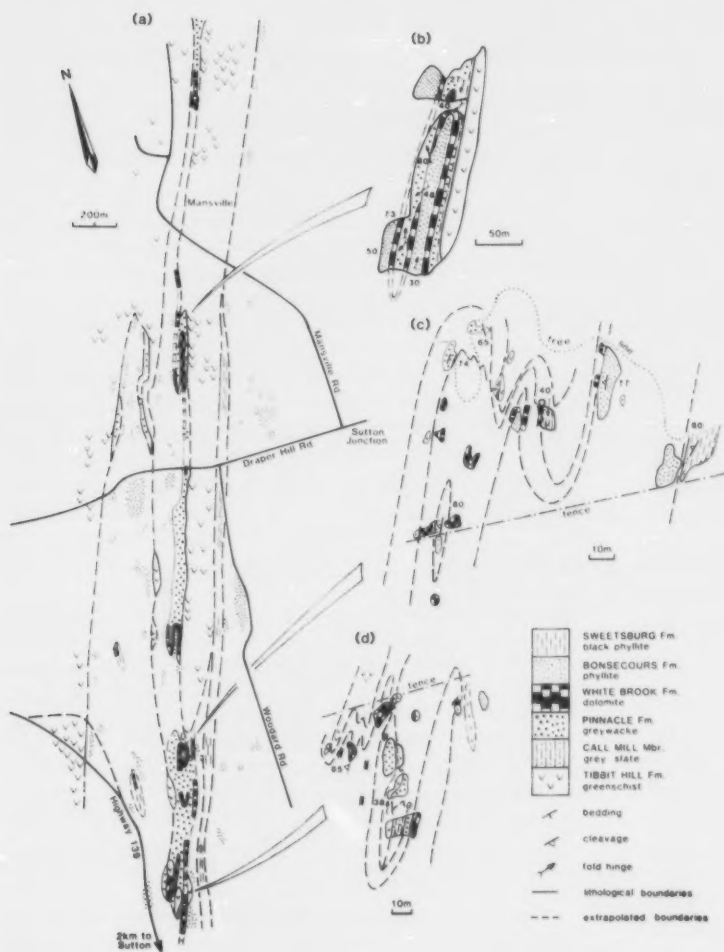


Figure 3. Outcrop of tightly folded Oak Hill formations of the Mansville facies, west of Sutton Junction. **a)** Simplified outcrop map of the main zone from Mansville to Sutton. **b)** Sketch of outcrop based on plane table map by T.H. Clark (pers. comm.). Note variation of fold plunge. **c)** Plane table map. Note shearing out of Pinnacle greywacke. **d)** Notebook sketch map. Note tight anticlinal cores of Tibbit Hill volcanics.

but much of the thinning is due to structural attenuation in the Brome Slide Zone. These problems are dealt with later.

Eastern Zone (Mansonville-Lake Memphremagog Complex)

The Eastern Zone in this report includes all of the rocks above the Bonsecours Formation on the eastern limb of the Sutton Anticline (Table 1). The "Memphremagog Complex" of Clark (1934, 1936b) comprises the Mansonville slate, Magog slate, Bolton igneous series and Glenbrooke Group. Later Cooke (1950) reassigned the first two units to the Caldwell and Beauceville groups. Dresser and Denis (1944) and de Römer (1957, 1958, 1961) used Caldwell in an older more restricted sense for the upper part of the Mansonville, and this corresponds to the Stowe Formation in Vermont (Cady, 1960). The lower part of the Mansonville was recognized as Sweetsburg by Osberg (1956, 1965) when he mapped the formation around the nose of the Sutton Anticline (Fig. 1) to correlate with the Ottawaquchee Formation in Vermont (Cady, 1960). The term Mansonville is therefore discarded here. St. Julien et al. (1972) redefined the Beauceville Formation of the Magog Group, and placed the lower shale-breccia-volcanic rocks assemblage in a newly defined St. Daniel Formation, which may be partly correlative with the Caldwell (see Table 1).

In this area it was possible to map units representing the Sweetsburg, Caldwell and St. Daniel formations, as well as several igneous units; for convenience, the latter are described together. A synclinally infolded outlier of the Silurian Glenbrooke Group was also mapped.

Sweetsburg Formation

The Sweetsburg Formation in the Eastern Zone comprises the distinctive black phyllites and grey quartzites that outcrop along the western side of the North Missisquoi valley. The unit is about 300-500 m thick and is well exposed in cuts along Highway 243. Abundant white to fawn or grey quartzite interbeds are as much as 10 m thick, thicker than in the type Sweetsburg in the Western Zone. The thickest units are indicated on the map. They vary from massive to finely laminated, have grey slaty partings and are interlayered in places with sandstone containing coarse (2-3 mm) quartz and feldspar grains. Most beds are pyritic and sericitic quartzites, and some contain carbonate. The less resistant black phyllites contain the thin quartzose laminae and abundant pyrite characteristic of the typical Sweetsburg in the west.

The basal contact with the Bonsecours Formation is sharp and easy to map. It is probably a tectonic break because lenses of serpentinite occur along and near the contact, but there is little outcrop evidence for the major fault mapped by Ambrose (1942) and Cooke (1950).

Caldwell Formation

The Caldwell Formation is only about 200 m thick in this area. It was not recognized by Lamothe (1979, 1981) who

included it with the St. Daniel Formation in the Bolton-Centre district and with the Sweetsburg Formation in the south. It is distinguished from the Sweetsburg by the presence of impure schistose quartzite and greywacke, some of it very fine grained, and by generally noncarbonaceous greenish-grey phyllitic interbeds. The quartzites range from about 1-3 m in thickness and contain black quartz grains, feldspar and slaty rock fragments set in a schistose matrix of sericite, chlorite and fine lithic fragments. Greywacke beds are more abundant to the north near Bolton-South (de Römer, 1960).

Metabasalt layers and lenses lie close to the top and bottom of the formation. Lamothe (1981) mapped them as the Potton Springs formation but did not define it. The nature of the upper and lower contacts of the Caldwell Formation is obscure because of inadequate exposure and the presence of the volcanic rocks. Lamothe (1981) mapped most of the boundaries as faults. The lowermost metabasalt layer may be equivalent to those in the Stowe Formation in Vermont, or it may be a tectonic intercalation of the overlying unit. The large ultramafic lens at the upper junction indicates an important tectonic break, although, to the south in Vermont, Cady (1960) has reported a transitional upper contact.

St. Daniel Formation

Most of the hills along the western side of Lake Memphremagog are composed of basic volcanic rocks of the Bolton Member, and the poorly exposed areas between the hills are underlain by black pyritic slates with thin (1-15 cm) interbeds of grey quartzitic phyllite and light to dark grey quartzite. These layers are commonly broken into fragments; some may well be sedimentary breccias but in many cases the shredding is related to the axial-plane cleavage of local folds (see Fig. 17e,g). A good exposure of such a tectonic breccia occurs 0.5 km south-east of the cemetery, 2.5 km south of Knowlton Landing.

It is difficult to calculate the thickness of this unit because of the folding and brecciation, but there are probably over 1000 m of slates. The age of the formation is uncertain. North of this area it is unconformably overlain by Middle Ordovician graptolitic shales of the Beauceville Formation (St. Julien and Hubert, 1975). The far eastern part of the slates southeast of Knowlton Landing may well belong to the Beauceville Formation, but no boundary could be mapped. The lower part of the formation is interpreted as being coeval with the Caldwell Formation and perhaps even older units (St. Julien and Hubert, 1975).

Igneous rocks in the Eastern Zone

Within the Cambro-Ordovician rocks of the Eastern Zone, ultramafic rocks, metabasalts and gabbros are arranged in three distinct belts: a lower series of small serpentinite pods near the top of the Bonsecours Formation, a main zone of ultramafics and metabasalts (ophiolites) in the Caldwell Formation and an irregular zone of folded metabasalts in the St. Daniel Formation. All of

these rocks were originally included in the Bolton Igneous Group (Clark and Fairbairn, 1936), but it is useful to follow Cooke (1950) and restrict the name Bolton to the upper unit. Little work was done on these rocks during the present study; they have been described by Lamothe (1979, 1981). The ultramafics to the south of this area have been described in detail by Chidester (1953, 1968) and Stanley et al. (1984), and the Brompton-Thetford ophiolites to the north have been described by Laurent (1975) and Williams and St. Julien (1982).

In the Mansonville area the igneous rocks consist of lenses of serpentinized peridotite together with sheets or dykes of gabbro, diabase and metabasalt. The ultramafic lenses appear to be crosscutting but lie mainly within or near the top of the Caldwell Formation, sandwiched between metabasalt layers. This is likely a tectonic arrangement, as the ophiolites lie structurally above the Caldwell Formation to the north (Laurent, 1975), and this zone has been interpreted to be a highly faulted obduction suture (Laurent, 1975; Stanley et al., 1984). The serpentinite pods in the lower formations (Bonsecours Formation and also the Bolton Glen formation to the west) have been interpreted to be tectonically emplaced dismembered parts of the main ophiolite body (Laurent, 1975; St. Julien and Hubert, 1975), but they are more likely structurally emplaced on early thrusts (Doolan et al., 1982; Stanley et al., 1984; Stanley and Ratcliffe, 1985). The serpentinites are variably altered to talc in sheets and veins, accompanied by minor asbestos veins; the margins are brecciated in places and cemented by dolomite. Talc was mined in the small bodies 2 km west-northwest of Highwater and 5 km north of Mansonville, and is still being mined at the Baker Mine 2 km west of Bolton-South (Lamothe, 1981).

The gabbros have retained their ophitic texture although the pyroxene has been altered to green and yellowish-green amphibole, and the plagioclase has been saussuritized to calcite and zoisite rimmed with sericite. These gabbros may be feeder dykes to basalt flows. The basalts are schistose and altered, and contain segregated chlorite and epidote veinlets; no pillows were observed.

Bolton Member. The Bolton Member consists mainly of two thick (500 m and 150 m) layers of massive greenish-grey basaltic lava flows that underlie the isolated hills along the western side of Lake Memphremagog. They are equivalent to the Coburn Hill volcanics of the Missisquoi Formation in Vermont (Doll et al., 1961; Cady, 1969). The original mineral textures are preserved in some places, although, like the gabbro, the pyroxene and feldspar are altered. In other places, fresh sodic plagioclase is intergrown with bands of amphibole and the matrix contains epidote, sphene and accessory apatite. Chlorite spots, presumably tuff fragments, and amygdules were observed locally. Inclusions of black slate are common at the flow margins and rarely in the interstices of pillows.

Spectacular pillow structures (up to 2 m across) are preserved (see Fig. 17h) and give younging directions in several places (see Map 1750A). The pillows indicate a

uniform steep dip parallel to the general bedding in the slates and are folded with it. Although the margins are schistose, there can be little doubt about the conformability of the contact in most places (Doolan et al., 1982), and Cooke's (1950) observations to the contrary were not confirmed (Fig. 17i).

Glenbrooke Group

Clark (1936b) and Boucot and Drapeau (1968) confirmed the existence of Silurian rocks at Lake Memphremagog. They named the 800 m-thick unit of the Glenbrooke Group and divided it into three formations. Lamothe (1981) reduced the lowest formation to member status. Only the southern part of the outcrop area was mapped in this study; the part north of Sargent Bay was taken from St. Julien (1965) and Boucot and Drapeau (1968). In this report and map only the basal conglomerate member is distinguished from the group.

The lower part of the group mapped here consists of fawn-grey and bluish-grey pyritic shales and slates with 7-10 cm interbeds of dark grey limestones spaced at a few centimetres to one metre intervals. These are rarely fossiliferous and a Dalmanites tail was found in a roadcut 1 km southwest of Vale Perkins. Some layers have shaly partings and others are cut by a well-developed slaty cleavage. The age of the Glenbrooke Formation is Late Silurian (Boucot and Drapeau, 1968).

The limestones of the Sargent Bay Formation occur mainly in the north and were not differentiated. The small downfaulted wedge of limestone at Mountain House Wharf 4.5 km north of Leadville is early Middle Devonian (Boucot and Drapeau, 1968).

Peasley Pond Member. Lenses of conglomerate as much as 30 m thick at the base of the Glenbrooke Group constitute the Peasley Pond Member. The matrix is a calcareous or gritty sandstone. Pebbles vary from 5-10 cm in diameter at the base reducing to 1 cm toward the top of the formation, and consist predominantly of rounded to subangular white vein quartz with lesser amounts of light grey, fine grained quartzite, black slate and volcanic rocks. The quartzite and slate are similar to those of the St. Daniel Formation. Volcanic fragments are rare even where the conglomerate rests directly upon volcanic rock, but some gabbro and basic volcanic pebbles are present and volcanic detritus occurs as gritty lenses in the matrix.

Granite

Coarse grained biotite-muscovite granite outcrops along the shore of Lake Memphremagog north of Leadville; it is probably a wide dyke. Immediately to the south pyrite, galena and sphalerite occur in a few 3 m-thick quartz veins that cut an altered dyke of granite; an old lead prospect is located there. The granite undoubtedly belongs to the suite of Middle Devonian (Acadian) plutons that outcrop on the eastern side of Lake Memphremagog.

STRATIGRAPHIC CORRELATION AND PALEOGEOGRAPHY

The general paleogeography of this region has been elucidated in terms of geosynclinal theory in a series of papers by Cady (1960, 1967, 1968b, 1969) and Osberg (1969), and in terms of plate tectonics by Bird and Dewey (1970), Laurent (1975), St. Julien and Hubert (1975), Osberg (1978) and St. Julien et al. (1983). In relation to these theories the important structures of this area are: the Richford-Valcourt Syncline, which marks the transitional boundary between the miogeoclinal Oak Hill Group and eugeoclinal Sutton schists; and the ultramafic belt near Mansonville, which represents an obducted ophiolite complex separating Cambro-Ordovician oceanic deposits in the east from continental margin deposits in the west.

Summary comments here are restricted to setting the Cowansville-Sutton-Mansonville area into this general picture. Stratigraphic correlation is considered with reference to Table 1 and the cross section; a summary diagram (see Fig. 4) sets out the interpreted relationships in a paleogeographic framework.

Stratigraphic correlation

Correlation from west to east across the Green Mountain-Sutton Mountain Anticlinorium depends on the demonstration that the Sweetsburg Formation passes across the Sutton Anticline into the Ottauquechee Formation (Osberg, 1956, 1965, 1969; Cady, 1969). A second parameter essential to the correlation of the Oak Hill Group with the Sutton schists depends on an interpretation of the Mansville facies. This latter aspect is dealt with first.

It is clear from regional considerations (Cady, 1969) that a major eastward thickening of sediments accompanied by a facies change occurs along the line of the Richford-Valcourt Syncline, and that parts of the Sutton schists are correlatives of Oak Hill sediments (Clark and Eakins, 1968). Osberg (1965) has traced the lower Gilman phyllite into the Bonsecours schist and, at a lower stratigraphic level, Cady (1960) and Dennis (1964) have shown that in Vermont the Underhill schist (= Bonsecours) passes transitionally westward into the Pinnacle greywacke and eastward into the Hazens Notch schist. Thus, there is an eastward thickening of the shale-silt facies. The thinning of the Oak Hill sediments in the

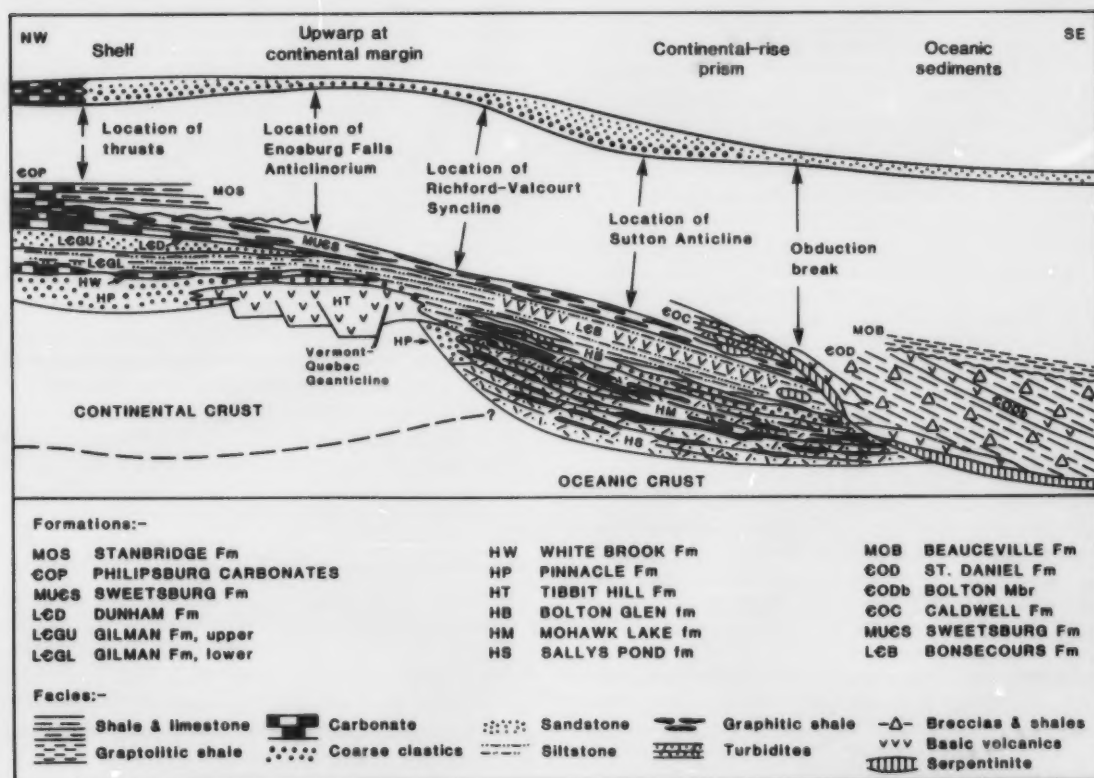


Figure 4. Reconstructed paleogeographic section for the Middle Ordovician continental margin prior to major tectogenesis. Note the interpreted correspondence between paleogeographic features and later major structures. The distance from northwest to southeast in the figure is on the order of 200 km.

Mansville facies was achieved by sedimentary and tectonic processes. The thinning is due partly to eastward convergence toward the Vermont-Quebec Geanticline (Cady, 1969), but it is difficult to explain how all of the units, especially the dolomite and ferricrete, thin so uniformly without local facies changes. It is true that the more massive beds of the Pinnacle Formation have not been involved in the tight folding of the Mansville facies but the greywackes occur close by and it is likely that they were too competent to become so tightly folded. Nevertheless, typical Pinnacle greywackes are not found east of the Mansville outcrops and the upper Gilman quartzite is not represented. The occurrence of Sweetsburg slate with thin Dunham dolomite within the zone of the Mansville facies indicates that the intervening lower Gilman phyllite and siltstone must also be thinned and yet these are the lithologies that become dominant in the major facies change. These observations support the case for both sedimentary and tectonic thinning; additional evidence for structural attenuation will be presented later.

The thinning in the Mansville facies makes correlation with the Sutton schists difficult. The interpretation advocated here is to correlate the Tibbit Hill volcanics with the Rosenberry greenschist and the Pinnacle greywacke with the quartzofeldspathic schist of the Bolton Glen formation (see cross section). The marble lenses in the Bolton Glen formation then correspond with the White Brook dolomite. According to this interpretation the Bonsecours Formation is equivalent to both parts of the Gilman Formation, and the Mohawk Lake and Sallys Pond formations lie beneath the Tibbit Hill volcanics (see cross section and Table 1).

The internal stratigraphy of the Sutton schists is presented as a simple sequence, as no evidence was found for the early nappes reported to the north (St. Julien and Hubert, 1975) and south (Stanley and Ratcliffe, 1985). Cady et al. (1963) demonstrated that the Hazens Notch (= Mohawk Lake) Formation underlies the Ottawa-Quebec Formation and intertongues with the Underhill Formation and perhaps the Pinnacle Formation, but more recent mapping in northern Vermont indicates that the sequential units of Underhill, Hazens Notch and Ottawa-Quebec are separated by thrusts (Doolan et al., 1982; Stanley and Ratcliffe, 1985). In the report area, the simple closure of the Sweetsburg and Bonsecours formations around the northern end of the Sutton Anticline (Osberg, 1965) precludes major nappe stacking, at least after the Cambrian, unless there is a major folded thrust at the base of the Bonsecours Formation.

The Gilman quartzite contains Early Cambrian fossils (Clark, 1936a) but these are not of earliest Cambrian age (Rodgers, 1972), so that the base of the Cambrian cannot be identified there. Clark (1934, 1936a) recognized that part of the Oak Hill sequence might be of Precambrian age. The boundary between the Cambrian and Precambrian in this report is arbitrarily placed at the postulated disconformity above the White Brook dolomite, a break that corresponds to the contact between two sedimentary cycles (Clark, 1936a). Thus all of the Sut-

ton schists below the Bonsecours Formation are considered to be Hadrynian (Table 1).

Higher in the sequence, the Bonsecours Formation has been traced across the nose of the Sutton Anticline, the only facies change being a possible increase in volcanic and carbonaceous layers. The Sweetsburg Formation also passes across the Sutton Anticline with only a slight change in the silty laminae to thicker sandstone beds eastward. The upper parts of the sequence in the east are more difficult to correlate, and St. Julien and Hubert (1975) assigned older ages than Cady (1969). The Caldwell Formation has been assigned an Early Cambrian age (St. Julien and Hubert, 1975; St. Julien et al., 1983) and a Cambrian to Middle Ordovician age (Doolan et al., 1982). There is no western equivalent to the Caldwell Formation in this area, unless it be the Vail Member (see Table 1); it was either not deposited, or was eroded before the Stanbridge Formation was deposited, or is buried beneath a possible thrust. In places on the Sutton Anticline, equivalent rocks have been tectonically removed to constitute the Granby nappe to the northwest of the report area (St. Julien and Hubert, 1975).

The St. Daniel Formation appears to overlie the Caldwell Formation, but the junction is a tectonic one, marked by serpentinites and volcanic rocks of an ophiolite suite. St. Julien and Hubert (1975) and St. Julien et al. (1983) have interpreted the St. Daniel Formation as an ensimatic facies equivalent to the entire Cambrian and Lower Ordovician sequence west of the serpentinite belt (Table 1).

The relationship between the Stanbridge complex and the eastern sequence is also unclear (Fig. 4), as much of the Stanbridge Formation is allochthonous (St. Julien and Hubert, 1975). The early Middle Ordovician age determined from graptolites in the Cowansville quarry (Riva, 1966) shows that the slates there are older than the Middle Ordovician Beauceville slates that lie unconformably above the St. Daniel immediately to the north of Sargent Bay (St. Julien, 1965). If, as suggested earlier, the Stanbridge and Sweetsburg formations are in sedimentary contact or only slightly disconformable, the sequence from Middle Cambrian to Middle Ordovician is very thin in the western part of this area.

Paleogeography

By way of summary of the stratigraphy and introduction to the structural section it is convenient at this stage to examine the local paleogeography immediately prior to deformation in the Middle Ordovician. This situation has been reconstructed from the geological section, from geophysical data and from the nature of the sediments and the stratigraphic correlations suggested above (Fig. 4). Major structures have been simply unravelled, but no attempt has been made to produce a balanced palinspastic reconstruction.

Geophysical data, mainly from the north of this area, have yielded important crustal models. Major Bouguer gravity and magnetic anomalies coincide with the Notre Dame and Sutton mountains (Innes and Argun-Weston,

1967; Seguin, 1982, 1983; Williams and Haworth, 1984a,b). Similar high anomalies occur along the length of the Appalachians (Cook and Oliver, 1981; Hatch, 1982; Williams and St. Julien, 1982) and are typical of fractured continental margins, similar to the present-day Atlantic margin (Rabinowitz and LaBrecque, 1977; Grow et al., 1979; Keen and Hyndman, 1979). Bird and Dewey (1970) placed the Taconic shelf margin at the western edge of this gravity high.

Deep seismic reflection profiles to the north indicate that the continental basement is step faulted and dips gently eastward to a position beneath the western edge of the schist outcrop in the Notre Dame Anticline (Laroche, 1983; Cook, 1984); it may terminate there (Seguin, 1983) or its continuation may be masked by seismic noise (Cook, 1984). The basement reflector traces farther east than the density contrast necessary to explain the gravity anomaly (Cook, 1984), and Kumarapeli et al. (1981) and Seguin (1983) explain the anomaly as due to a wedge of Tibbit Hill volcanics in a continental-margin rift extending down to the basement reflector at about 10 km. Cook (1984) argues for an additional deeper dense source, and he postulates a transitional continent-ocean crust boundary buried beneath an allochthonous slab containing basic volcanic rocks.

Kumarapeli et al. (1981) have shown that the peak of the gravity and magnetic anomalies coincides with the outcrop of the Tibbit Hill volcanics. Thus in the Sutton region the high anomalies correspond to the Enosburg Falls Anticlinorium, and the facies changes and intense deformation in the Richford-Valcourt Syncline lie immediately to the east, marking the boundary between the western and central stratigraphic zones.

In northwestern Vermont the pre-Cheshire clastic sediments form a coherent sequence typical of deposition in ensialic rift basins (Rodgers, 1972; Stanley and Ratcliffe, 1985). The lower part of the western Oak Hill sequence fits this pattern well, and the Tibbit Hill volcanics have been interpreted to be the product of rift-related volcanism during the early stages of ocean opening (Rankin, 1976; Kumarapeli et al., 1981; Coish et al., 1985; Pintson et al., 1985). The repetition of the shallowing-upward quartzite-dolomite deposition in the Oak Hill sequence, recognized by Clark (1936a), may have resulted from the development of thermal-sag basins over the initially rifted continental margin (compare Hoffman et al., 1974; Bott, 1980). After the rifts filled with basic volcanic rocks and greywacke, the Gilman and Dunham formations (shales, silts, sandstones, dolomite) and the Sweetsburg Formation (shales) were deposited and may represent the deposits of sag basins.

The lithology, chemistry and well-bedded nature of the Sallys Pond formation indicate that it was originally arkosic sand and shale (Table 2, cf. columns 4 and C); the alkali-silica ratio plots in the passive-margin field (Roser and Korsch, 1986). Thus the chemistry supports the interpretation that the Sutton schists were late Precambrian, as sediments of this age along the Appalachians are characterized by arkosic composition (Rodgers, 1972). The Mohawk Lake sediments were car-

bonaceous mud and impure sand probably deposited as turbidites. The Gilman, Pinnacle and Bonsecours formations have compositions close to an average shale or greywacke (Table 2, cf. columns 1, 2 and 3 with A and B) and they all plot in the active-margin field of an alkali-silica diagram (Roser and Korsch, 1986). These features and the presence of basic metavolcanics and serpentinites suggest that the Sutton schists may be interpreted as the oceanward eugeoclinal or marginal basin equivalents of the lower Oak Hill Group and older sediments (Table 1) deposited as a continental-rise prism at the rifted margin similar to the present Atlantic margin (Dewey and Bird, 1970; Rodgers, 1972). The hypothesis that the Sutton schists represent a continental-rise prism implies that the schists cover the continent-ocean crust boundary or thinned continental crust; a similar explanation has been proposed for the Bennett schists in the Notre Dame Anticline to the north (Seguin, 1983). Thus this boundary is inferred to be about 20 km west of the position of the ophiolite belt postulated by Laurent (1975), St. Julien and Hubert (1975) and St. Julien et al. (1983). This has important tectonic implications, as will be shown later.

During the Hadrnyian and Early Cambrian sedimentation, a geanticline, the Vermont-Quebec Geanticline (Cady, 1967, 1969), developed along the line of the present Enosburg Falls Anticlinorium (Fig. 4). This is manifested by the southward termination of the Tibbit Hill volcanics, the erosional disconformity above the White Brook dolomite and the exposure of older Precambrian rocks along the line of the Pinnacle Formation outcrop (Doll et al., 1961). The geanticline may have been a shelf-edge upwarp like that occurring at the present west Atlantic Ocean margin (Drake et al., 1961; Rabinowitz and LaBrecque, 1977). A similar situation is reported for the Newfoundland Appalachians (Williams and Stevens, 1974).

This geanticline continued to separate miogeoclinal and eugeoclinal facies throughout most of the Cambrian and Ordovician, although the facies vary obliquely across it (Cady, 1967, 1969; Osberg, 1969). During the Early Cambrian a marked upward fining of the sediments occurred as the Gilman and Bonsecours muds and silty sands were deposited. The upper Gilman quartzites and Dunham carbonates pinched out on the reduced (possibly buried) geanticline and by Middle Cambrian times carbonaceous muds with only thin sandy layers (Sweetsburg Formation) spread across the whole outer shelf and continental rise (Fig. 4). Extensive carbonate banks then developed in the west with a bank edge along the line of the present Lake Champlain (Rodgers, 1968).

A sudden change in the paleogeography occurred in the Early Ordovician when the "Atlantic-type" margin changed with the development of a subduction zone (Laurent, 1975; St. Julien and Hubert, 1975). The first effect was an influx of coarser sediments—the greywackes and shales of the Caldwell and Granby groups. Then ophiolites were obducted onto and possibly into the continental-rise sediments, and shale melange (St. Daniel Formation) was deposited as an island arc developed

(Williams and St. Julien, 1982). The latter is represented by the calc-alkaline volcanic rocks of the Ascot and Weedon groups to the east of this area (St. Julien and Hubert, 1975; Doolan et al., 1982). These volcanic rocks were developed on continental crust (Osberg, 1978) and an external strip of basement rocks has been recognized along the length of the Appalachians (Hatcher, 1983).

The ophiolite belt has been interpreted to be a suture representing the closure by subduction and obduction of the Iapetus Ocean (Bird and Dewey, 1970); it can be traced northward to Newfoundland as the Bai-Verte-Brompton line (Williams and St. Julien, 1982). The line cannot be traced southward beyond northern Vermont, which prompted Cady (1972) and Chidester and Cady (1972) to object strongly to the interpretation of the serpentinites as an oceanic suture; they argued that in Vermont, at least, the serpentinites are protrusions through a sialic crust. However, the ophiolitic association of the serpentinites in southern Quebec (Laurent, 1975) precludes an intrusive origin (Stevens et al., 1972). Moreover, recent geochemical studies indicate that rocks along this zone have different oceanic affinities (Clague et al., 1981; Shaw and Wasserburg, 1984; Coish et al., 1986) and the Thetford Mines ophiolites comprise island arc volcanic rocks overlying oceanic basalts (Laurent and Hébert, 1978). Detailed remapping of the northern Vermont and southern Quebec outcrops demonstrates extensive faulting consistent with an accretionary margin, and the ophiolites have been dismembered by strike-slip faults and structurally telescoped (Doolan et al., 1982; Stanley et al., 1984; Coish et al., 1986). A marginal ocean basin may have terminated southward as advocated by Cady (1972), and Hiscott (1984) concluded from a study of trace elements (especially Cr and Ni) in Ordovician flysch sediments that ophiolite obduction did not occur or was of minor importance in the United States Appalachians. More likely the suture may be buried by a thrust sheet (Williams and St. Julien, 1982; Stanley and Ratcliffe, 1985). Ordovician paleogeography was evidently one of festoons of microcontinental and volcanic arcs (Osberg, 1978; Williams and Hatcher, 1983).

The small serpentinite lenses in the Bolton Glen formation and between the Bonsecours and Sweetsburg formations (Fig. 4) may mark thrusts (Doolan et al., 1982; Stanley et al., 1984) possibly splaying off the main serpentinite zone (St. Julien and Hubert, 1975), or representing earlier aborted obduction slices formed by periodic detachments from the oceanic crust (Bird and Dewey, 1970). In this connection it is interesting to note that Trzciński (1976) reported the occurrence of crossitic amphibole in the Tibbit Hill volcanics in the Richmond area and postulated a subduction zone in this position west of the main ophiolite belt of St. Julien and Hubert (1975).

Tectonic activity probably started in the Early Ordovician and followed the course outlined by Rickard (1965) and St. Julien and Hubert (1975). Nappe slices of Cambro-Ordovician rocks slid or were pushed westward toward the foreland into muds that by then "mantled" the depressed shelf region (Zen, 1972; Stanley and Rat-

cliffe, 1985); the muds are probably represented by parts of the Stanbridge Formation. Ophiolites were obducted and thrust over and into the underlying Caldwell sediments, and the ocean floor shale and breccias of the St. Daniel Formation were telescoped against what is interpreted here as the collapsing Precambrian-Cambrian continental-rise prism (Fig. 4). The Middle Ordovician Beauceville muds were then deposited over the already deformed Caldwell Formation and were themselves deformed (St. Julien, 1967).

As orogeny proceeded, the location of the major structures was controlled by the paleogeography (Fig. 4). The Enosburg Falls Anticlinorium developed more or less along the continent-edge geanticline, the Sutton Mountain Anticlinorium developed under rising metamorphic conditions in the collapsing continental-rise prism and later large thrust faults developed along the edge of the carbonate banks to the west (Rodgers, 1968). Siluro-Devonian sediments spread westward over the eroding Ordovician (Taconic) orogen and were folded during later orogenic activity (Acadian) to form the Sargent Bay Syncline.

In the following sections, the structure and deformation history are described in detail.

STRUCTURAL GEOLOGY

Five major structures have been mapped (Fig. 1 and 5, in pocket). These are, from west to east:

1. *Enosburg Falls Anticlinorium*—the western limb was overturned in the Oak Hill and Hunter Mills anticlines and thrust westward toward the foreland sequence. The crest consists of a zone of tight upright folds, which are well defined by the formations of the Oak Hill Group (see Map 1750A).
2. *Richford-Valcourt Syncline*—a tight major fold at the junction between the Oak Hill Group and the Sutton schists is characterized by a strong attenuation of strata and steeply-plunging to reclined* minor folds.
3. *Sutton Anticline*—a broad simple arching that folds the schistosity and early east-west reclined folds in the central schist belt.
4. *Eastern Complex*—a zone of large steeply-plunging reclined folds, including Sugarloaf and Bear Mountain folds.
5. *Sargent Bay Syncline*—a tight, gently-plunging and upright fold of Silurian rocks set within the reclined folds of the Eastern Complex.

Two sets of folds are apparent over most of the area but the geometry of early minor structures in each of the five zones is sufficiently different to warrant separate description. The late folds, however, have a more uniform geometry and all of the zones are treated together. The geometry, possible correlations and modes of development are then summarized.

* A reclined fold plunges down the dip of its axial plane, i.e. its hinge has $90^\circ \pm 10^\circ$ pitch (Fleuty, 1964a)

Structural elements

The planar elements mapped consist of bedding, bedding schistosity, axial-plane slaty cleavage and schistosity and crenulation cleavage. Lineations comprise fold hinges, axes of crenulations, intersections of planes, quartz rods and mineral alignment orientations ("grain"); except for the last, most of the lineations are clearly related to axes of folds and were assigned to early or late categories. The "vergence" of folds was recorded wherever possible. A selection of measurements is plotted on a structural map (Fig. 5, in pocket) and all of the data are plotted on stereographic projection diagrams (Fig. 6, in pocket); early and late structures are distinguished. The area was divided into 26 domains for the purpose of structural analysis, and these are grouped along the main structures and presented in approximate relative geographic posi-

tion (Fig. 6). A profile of the cleavage relationships is included in the cross section. The relationships among the structural elements in the different situations are arranged to give a realistic illustration of structural styles in the area (Fig. 7).

Early structures

Enosburg Falls Anticlinorium

The name Enosburg Falls Anticlinorium is taken from Doll et al. (1961). Structures related to the anticlinorium control the outcrop pattern of the Oak Hill Group. The attitude of minor folds varies from upright in the crest to overturned on the western limb and the well-developed axial-plane foliation fans markedly (see cross section).

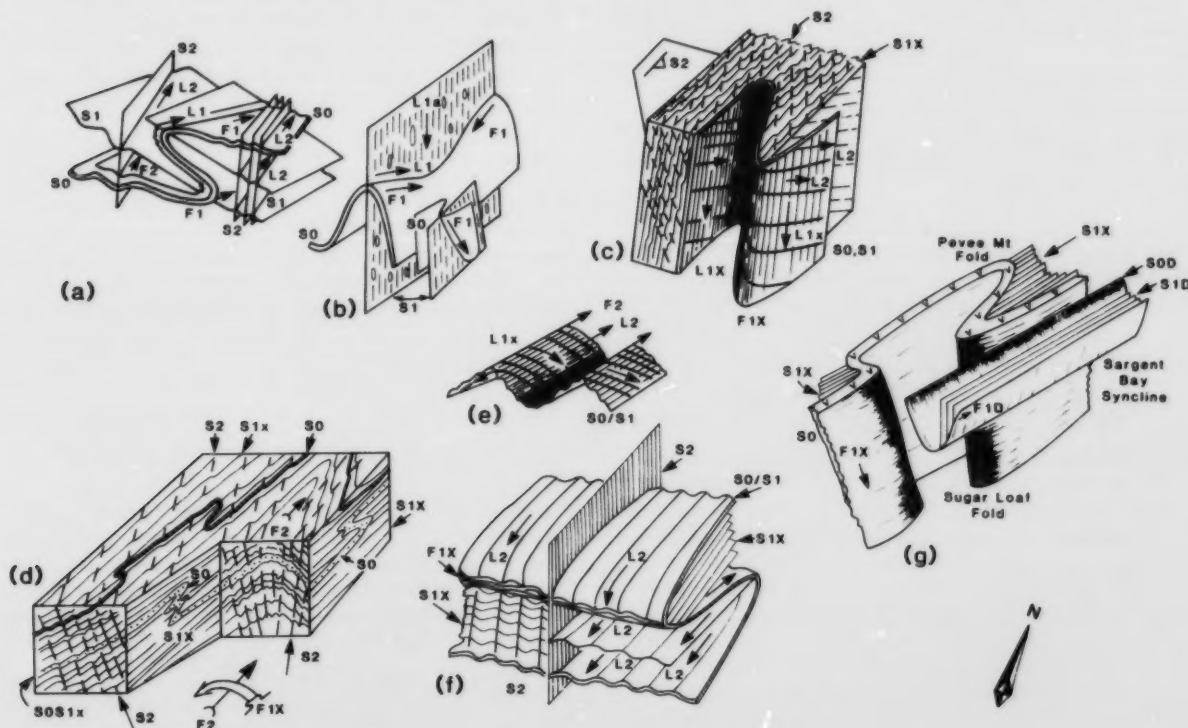


Figure 7. General arrangement of structural elements across the area. Bedding (S_0) is folded into early north-south folds (F_1) and east-west cross folds (F_{1X}); axial-plane cleavages S_1 and S_{1X} intersect bedding to form lineations L_1 and L_{1X} , respectively. A late crenulation cleavage (S_2) is axial planar to late folds (F_2) and crenulation lineation (L_2). **a)** Represents the structure of Oak Hill and Hunter Mills anticlines in the west. Overturned to recumbent folds (F_1), with axial-plane slaty cleavage (S_1) are refolded by upright late folds (F_2) with crenulation cleavage (S_2). **b)** Represents the structure at the crest of Enosburg Falls Anticlinorium with variably-plunging folds (F_1) with steeper folds in incompetent layers. Axial-plane slaty cleavage (S_1), with mineral elongation or stretching lineations (L_{1a}). **c)** Represents the structure in the Richford-Valcourt Syncline. Here, steeply-plunging tight folds (F_{1X}) with axial-plane crenulation cleavage (S_{1X}) are cut by late crenulation cleavage (S_2). **d), e)** and **f)** Represent structures in the Sutton Anticline. Bedding schistosity (S_0/S_1) is folded by east-west reclined cross folds (F_{1X}) which also have quartz-rod lineations (L_{1X}) and an axial-plane crenulation cleavage (S_{1X}). These elements are all folded into a major anticline (F_2) with an upright crenulation cleavage (S_2). Where the S_2 crenulation cleavage cuts across the nose of an early fold (F_{1X}) then the late crenulation lineation (L_2) is steep (**f**). **g)** Represents the structure of the Eastern Zone. Large steep reclined folds related to cross folds (F_{1X}) with axial-plane slaty cleavage (S_{1X}) are overlain unconformably by Siluro-Devonian rocks folded into a gently-plunging syncline F_{1D} with axial-plane slaty cleavage (S_{1D}).

Oak Hill and Hunter Mills anticlines

The upper Gilman quartzite and overlying formations dip steeply west, or are overturned to the west, from Sweetsburg to the International Boundary southwest of Frelighsburg (map and Fig. 10). These steep beds delimit the western limbs of two folds (Fig. 5). The eastern limbs are more gently dipping so that the folds are asymmetrical (Fig. 8a,b,c, 9d and 10). A moderately east-dipping slaty cleavage marks the axial planes of the folds and this cleavage is commonly emphasized by wafer-thin quartz veinlets (Fig. 8e), although as noted by Williams (1966) the quartz is later and cuts across the cleavage in places (Fig. 8f).

This structural pattern is found along the whole zone (Fig. 6, domains 1 to 3). Minor fold hinges and bedding-cleavage intersections plunge gently to the northeast or southwest spreading slightly in the girdle of the cleavage plane; bedding poles plot in incomplete broad girdles about the main northeast-southwest lineation clusters. Charbonneau (1975b) has reported a similar pattern from the Frelighsburg area, but he attributed the spread of lineations to warping across earlier folds; as explained later, this variation is considered here to be an integral feature of early fold geometry.

The variation in fold style is strongly controlled by lithology as reported by Clark and Eakins (1968) and is illustrated in Figure 8. Slaty cleavage develops only in the pelitic layers. The massive Dunham dolomite has a rough

wide-spaced cleavage, and remnants of bedding may be traced across it in places (Fig. 8f). The upper Gilman quartzite is shredded into lenses, with slaty pelitic layers penetrating along the cleavage. This structure is most marked in the Gilman phyllites where shredding was so severe that silty layers were disrupted and transposed into the cleavage plane (Fig. 8d).

Insufficient outcrop makes observations on the nature of the contact between the Vail and the Stanbridge slates difficult, but there is no evidence, such as mylonites or breccias, to indicate a structural discordance. Structurally the Stanbridge slates are little different from the Vail slates except that early folds are tighter, are recumbent and disrupted and lack obvious axial-plane slaty cleavage (Fig. 9); they may be sedimentary slump folds. Williams (1966) made a detailed study of the Rioux Quarry at Cowansville; the pattern of structural elements there is similar to that in the adjacent Oak Hill rocks, except that the trend of early fold hinges varies more widely and bedding planes are buckled by late folds (cf. Fig. 9 and Fig. 6, domains 1 to 3).

Hinesburg Thrust

Clark's (1934, 1936a) postulated Oak Hill Thrust is thought to be merely the zone of overturning along the western limb of the Oak Hill Anticline (Fig. 10). However, the Oak Hill and Hunter Mills anticlines are separated along Leavitt Creek by what is interpreted as a

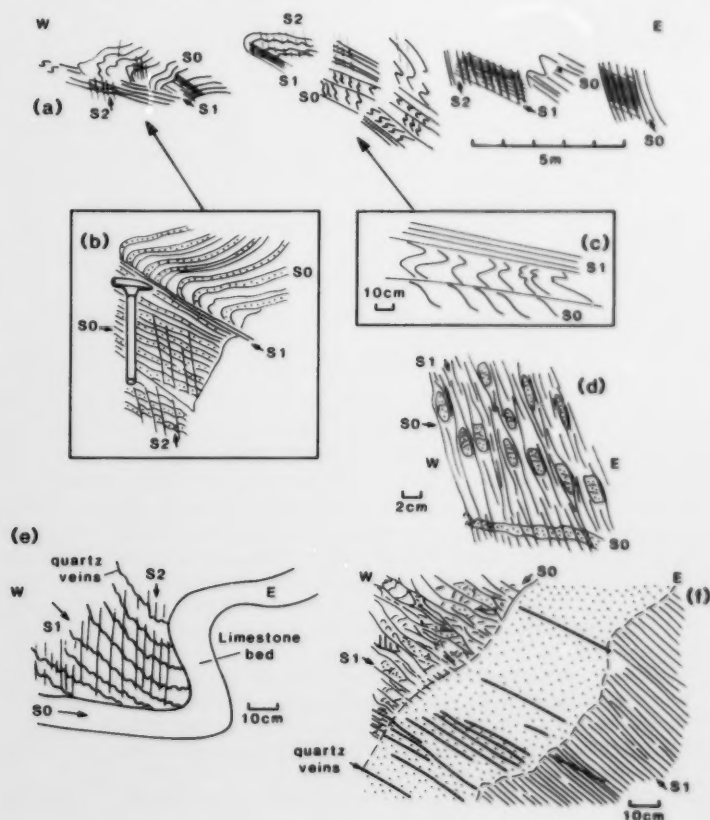


Figure 8. Early F_1 structures in the upper part of the Oak Hill Group. **a)** Cross section through early recumbent folds in the Vail Slate. Location—roadcut 3 km northeast of Dunham (745024). **b)** and **c)** Details from **a)** showing the relationship of minor thrusts to early slaty cleavage (S_1) and late crenulation cleavage (S_2). **d)** Shredding of siltstone layers in the lower Gilman phyllite. Location—1.3 km east-southeast of Oak Hill (787036). **e)** Quartz veinlets in slaty cleavage (S_1), cut by late crenulation cleavage (S_2). Stanbridge Formation. Location—Rioux Quarry, Cowansville (782075). **f)** Bedding and early cleavage (S_1) in Dunham dolomite. Quartz veins parallel and cut across cleavage (S_1). Location—4 km northeast of Dunham (762026).

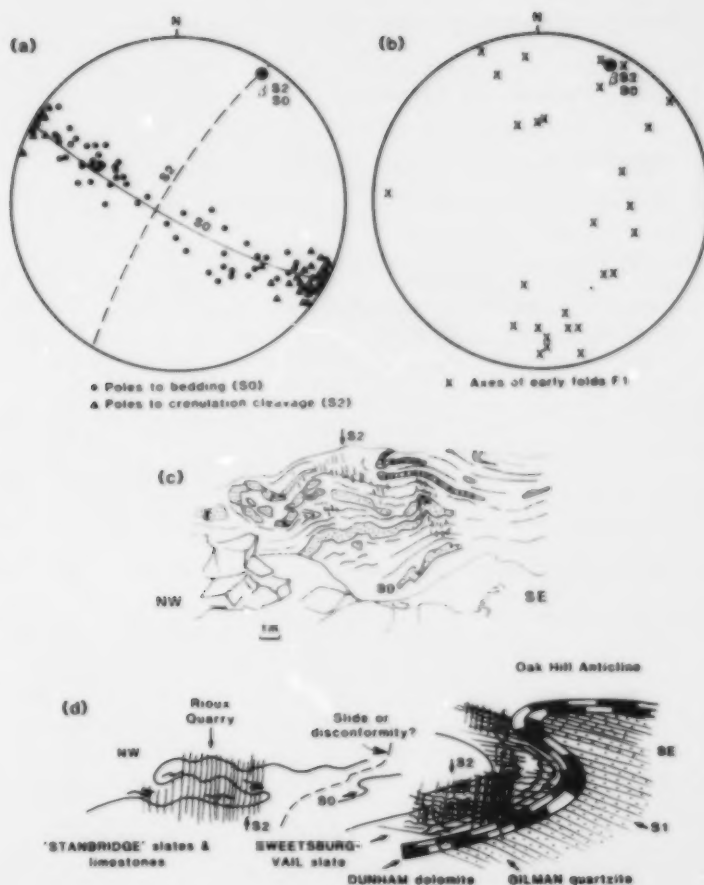


Figure 9. Refolded folds in the Stanbridge Formation in the Rioux Quarry, Cowansville (782075). **a)** Stereogram of poles to bedding (S_0) and late crenulation cleavage (S_2). **b)** Stereogram of early fold axes (F_1). **a)** and **b)** from Williams (1966), with permission. **c)** Refolded early fold; limestone bed (dotted) in black shale. **d)** Sketch section showing the relationship of Stanbridge structures to the Oak Hill Anticline.

continuation of the major Hinesburg Thrust (Dennis, 1964) (Fig. 1 and 10). Local faults at steeply overturned beds and some fault breccia support the thrust hypothesis (Fig. 10). A section across the fault zone (Fig. 10) indicates a normal (lag) displacement rather than overthrusting; this is attributed to local variations in fold plunge and the probable northward termination of the thrust. In northern Vermont the thrust has a maximum displacement of 10 km with a northwesterly slipline; it dies out southward into the attenuated lower limb of a recumbent fold (Stanley and Ratcliffe, 1985).

Pinnacle, Call Mill and Sweet Mine synclines

A series of synclinal outliers of Pinnacle greywacke and younger rocks occur in the Tibbit Hill volcanics at the crest of the Enosburg Falls Anticlinorium. This zone is characterized by upright, open to tight folds with a strong axial-plane slaty cleavage (Fig. 11 and 12). There is a strong lithological influence on fold style, and the competent greywackes control the overall structure, forming a variable but gently-plunging array of synclines and anticlines on all scales (cf. Fig. 11a and map). The detrital magnetite layers and quartzitic layers form near-parallel folds with little or no cleavage even where very tight (Fig. 11f and 12i). Massive greywacke beds have a rough axial-plane foliation; more pelitic beds have a good

slaty cleavage, as do the West Sutton and Call Mill slates and Tibbit Hill volcanics (Fig. 11b,d,e and 12a,b,g). Folds in these incompetent beds are more flattened in the limbs, and bedding is partially transposed, so that zones of fold hinges are separated by zones parallel to the axial-plane foliation; these structures vary from 0.5-30 cm wide (Fig. 11c,e and 12a,b,e,f,h). The hinges of these folds have a generally steeper plunge and in several places differ abruptly from adjacent folds in greywacke (Fig. 7b). In a few places, these steeper folds have a crenulation cleavage separate from the late crenulation cleavage (Fig. 13a,b); this indicates that there may be two early foliations, a possibility that will be examined later.

In many places the cleavage has a visible "grain", a penetrative but fine mineral alignment lineation subperpendicular to the main folds; tuffaceous slates show this structure well (Fig. 12d). This lineation (L_{1a}) represents the extension direction in the cleavage plane.

Quartz occurs as irregular "knots" and veinlets in zones, especially at the contacts between different lithologies. Quartz veinlets are commonly subparallel to bedding and are folded or boudinaged (Fig. 12c) but in some places are markedly oblique to or follow the axial-plane cleavage. Where they are folded, quartz veinlets in hinges commonly have a steep plunge. The development of quartz veins is described in more detail in the chapter on Metamorphism.

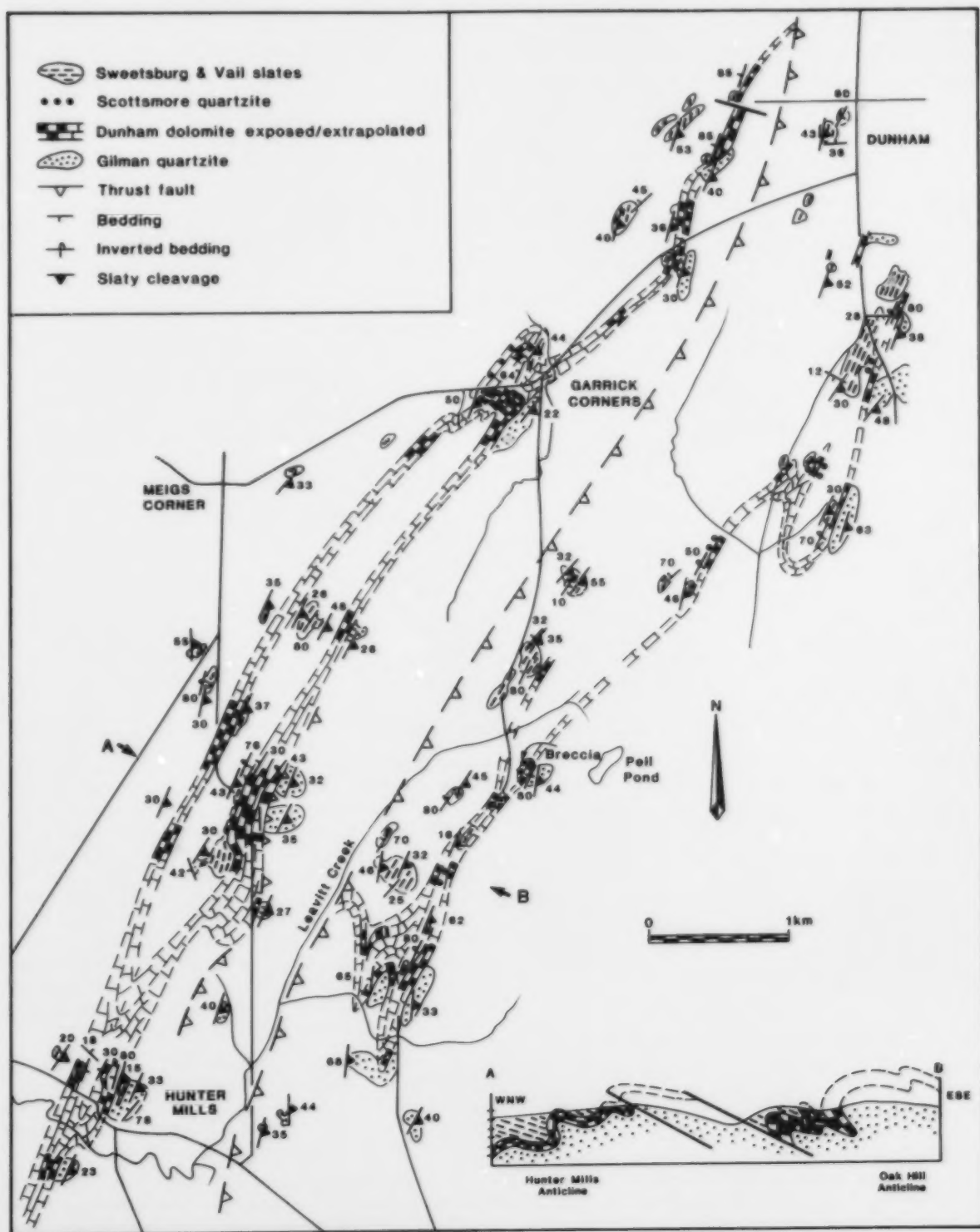


Figure 10. Hinesburg Thrust—outcrop map. Outcrop of upper Oak Hill formations southwest of Dunham along the line of the postulated Hinesburg Thrust.

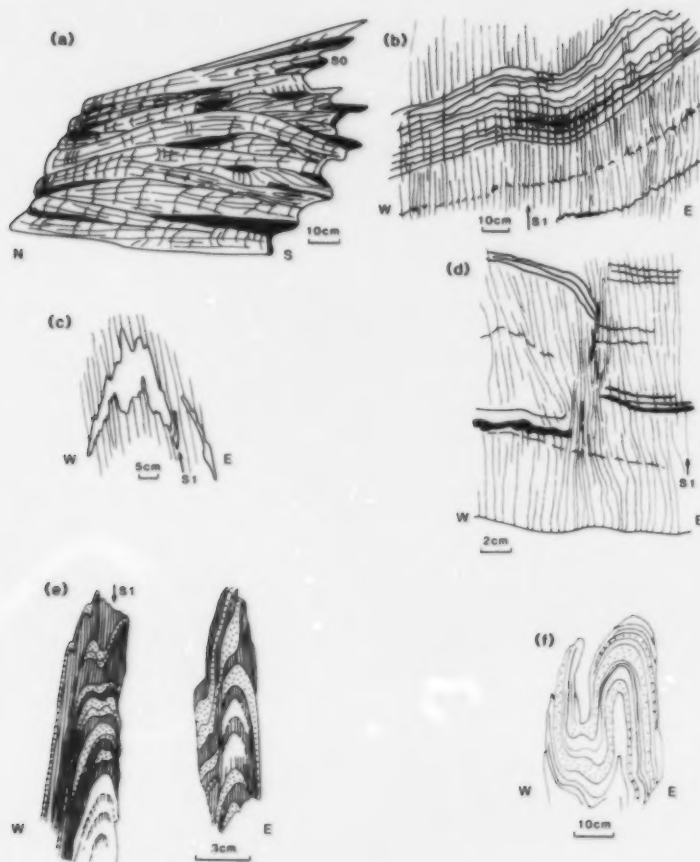


Figure 11. Style of typical early minor structures in the Pinnacle greywacke. **a)** Variably-plunging array of F_1 folds in detrital-rich layers. Location—1 km north of West Sutton (831991). **b)** Folded and cleaved thin detrital-rich layers. Location—1 km north of West Sutton (831991). **c)** Minor discordant fold in slaty layers. Location—West Sutton, old schoolhouse (837986). **d)** Disrupted detrital layers and slaty-type cleavage (S_1). Location—1 km north of West Sutton (831991). **e)** Strongly flattened folds in roadcut; Pinnacle greywacke of the Mansville facies. Location—Pinnacle Road, 2 km west of Alva (834923). **f)** Isoclinal fold in thin-bedded detrital-rich Pinnacle greywacke. Location—1 km south of Mansville (876025).

Structural analysis (Fig. 6, domains 5b to 9b) shows that the generally steep bedding planes are folded about north-northeast to south-southwest axes; the variability of fold plunge is shown by plots of minor folds and bedding-cleavage intersections, and by the broadness of the partial girdles of poles to bedding planes (see especially Fig. 6, domains 7b and 8b). The axial-plane cleavage dips steeply to the southeast or northwest (Fig. 6, domains 5c to 9c) and the "extension" lineation (L_{1a}) plunges more or less directly down dip. There is some indication of warping by late folding in domains 5c, 6c, 10c and 11c (Fig. 6). It is suggested later in this report that the spread of early-fold hinges (Fig. 6, domains 5b to 9b) is caused by a combination of initial plunge variation and by the rotation of hinges in incompetent beds toward the extension direction in the cleavage plane.

The structure of the lower Gilman phyllite is difficult to determine in detail because of insufficient and poor quality outcrops. Bedding probably has an essentially horizontal enveloping surface (see cross section) but it is locally steep and shredded into the steep southeast-dipping cleavage (Fig. 8d). Analysis of the meager data indicates that the geometry is similar to that of the rest of the Oak Hill rocks (Fig. 6, domain 4b,c).

Richford-Valcourt Syncline

The Richford-Valcourt Syncline can be traced from Richmond (north of the report area) southward through the Sutton valley and far into Vermont over a total length of 120 km. Dennis (1964) named it the Richford Syncline. In Quebec it has been called the Sutton Syncline (Eakins, 1964; Clark and Eakins, 1968), Brome Syncline (Rickard, 1965) and Valcourt Syncline (Osberg, 1965). The name Sutton is rejected because it suggests an association with the Sutton Anticline which is a later structure, and the combined name is preferred.

The fold is defined mainly by the series of narrow outcrops of Sweetsburg Formation locally associated with Dunham dolomite within the Bonsecours schist (Dennis, 1964; Osberg, 1965). Structurally it is difficult to define except as a fan axis between the steep foliation of the Enosburg Falls Anticlinorium and the west-dipping S_1 schistosity of the Sutton Anticline (see cross section). The syncline is overturned toward the east and more strongly so in the Richmond area to the north (Osberg, 1965; Marquis, 1985). The alternation of narrow zones of Sweetsburg and Bonsecours strata in the core zone near Abercorn probably marks early subsidiary folds in the core of the syncline. The synclinal form cannot be determined by vergence of minor folds, however (see below).

Figure 12. Style of typical early minor structures in the Tibbit Hill volcanics and Call Mill slate. **a)** Disrupted quartz-sericite layers (black) in grey slate. Location—1 km east-southeast of Call Mill (881055). **b)** Quartz-sericite phyllite (dotted) in grey slate near the contact with Tibbit Hill volcanics; early slaty cleavage (S_1) is cut by rough crenulation cleavage (S_2). Location—2 km west of Brome (885067). **c)** Feldspathic pod with vein quartz (black) in chlorite schist; schistosity (S_1) is cut by rough crenulation cleavage (S_2). Location—2 km north of Call Mill (874080). **d)** Talcose "willow-leaf" tuff fragments in slaty Tibbit Hill volcanics. Tuff fragments are oriented parallel to the mineral "elongation" lineation (L_{1a}) and crenulated by two later lineations (L_{1x} and L_2). Location—Highway 139, 3 km north-northwest of Sutton (864005). **e)** Local development of S_1 schistosity as a crenulation cleavage in thinly laminated and amygdaloidal Tibbit Hill volcanics. Location—1.3 km east-northeast of West Sutton (844986). **f)** Early fold in laminated chlorite-feldspar schist. Location—0.7 km southeast of Call Mill (876053). **g)** Steep quartzite beds (S_0) in grey slate; slaty cleavage (S_1) is crossed by a close-spaced crenulation cleavage (S_2). Location—Call Mill (872058). **h)** Dark grey slaty bands (black) in S_1 crenulation cleavage. Original bedding S_0 shown as folded dark and light grey slate layers. Late crenulation cleavage (S_2) cuts across S_1 . Location—Call Mill (871057). **i)** Folded (F_1) semipelitic layers in chlorite schist of the Tibbit Hill volcanics. Location—1 km northwest of Brome (903079).

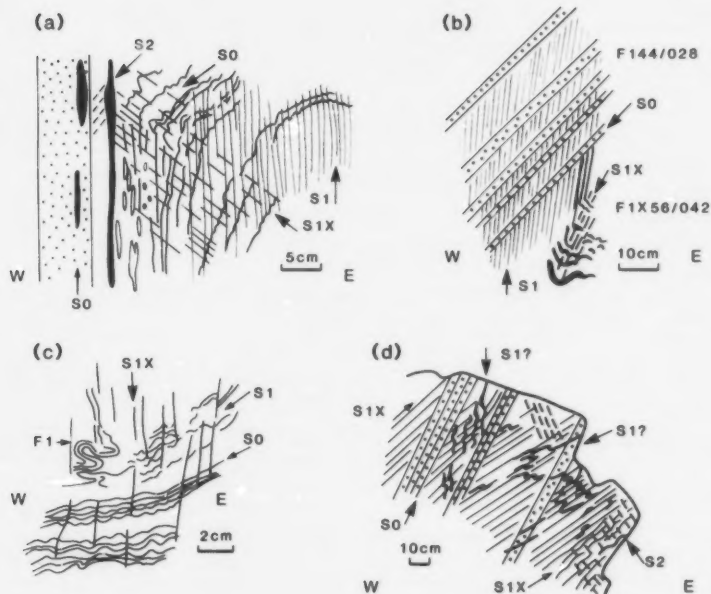
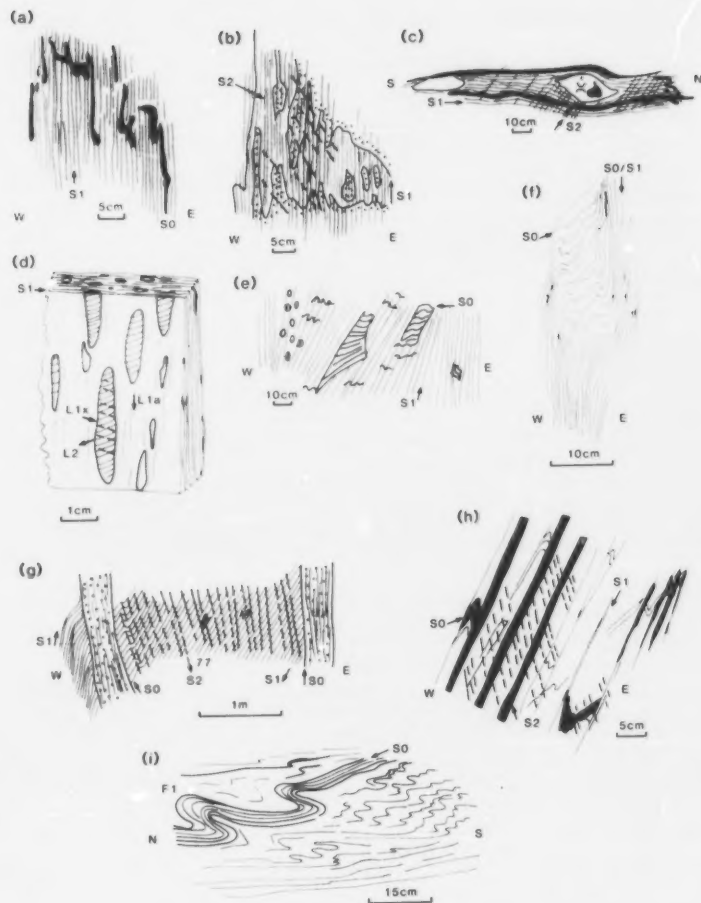


Figure 13. Possible superposition of two early foliations. **a)** A quartzite bed (dotted) and light and dark grey slate (S_0) is cut by slaty cleavage (S_1) and two crenulation cleavages interpreted to be S_{1X} and S_2 . Location—Call Mill (871056). **b)** Thin-bedded Pinnacle greywacke (dotted) and slate with folded quartz veinlets (black) in S_1 slaty cleavage are cut by a west-dipping crenulation cleavage (S_{1X}). F_1 and F_{1X} have different plunges. Location—3.5 km east-southeast of East Dunham (806971). **c)** Double fold relationships in the Bonsecours schist. F_1 predates the S_{1X} crenulation cleavage. Location—Alva (854921). **d)** In the Bonsecours phyllite, thin quartzite layers (dotted) represent bedding (S_0) and folded quartz veinlets possibly mark the S_1 cleavage; crenulation cleavage S_{1X} is axial planar to the quartz folds and is itself cut by a late crenulation cleavage (S_2). Location—2.5 km south-southwest of Brome (900046).

Bedding in the Bonsecours phyllitic schist is represented by thin quartzite laminae, which are commonly accentuated by parallel wafer-thin veinlets of quartz. The fine grain size and thinness of layers suggests strong attenuation, but any possibly original phyllonitic or mylonitic texture has been destroyed by later metamorphism. There are commonly two, and in places three, fine schistositys: the bedding schistosity and two crenulation cleavages (Fig. 7c, 13d and 15a). The earlier crenulation cleavage is associated with steep folds in the bedding schistosity and the second is the late foliation with a gently-plunging intersection with the bedding schistosity (Fig. 7c). Most quartz veinlets are parallel to the bedding schistosity, but some were probably crosscutting and subsequently deformed to give aberrant folds. The early crenulation cleavage is so strong in places that it forms a near perfect new schistosity; elsewhere mica and quartz have segregated to give a cleavage-parallel layering up to 2 cm wide (see Fig. 28). This foliation dips steeply throughout the syncline. It is plotted as S_1 on Figure 6 (domains 9c to 13c) as it is parallel to the upright cleavage (S_1) in the Enosburg Falls Anticlinorium, but it is suggested below that the foliation may be later, equivalent to the S_{1X} schistosity in the Sutton Anticline. The "elongation" lineation becomes difficult to recognize where the foliation is a crenulation cleavage, but in a few places steep fine "rib and groove" structures on quartz veins have been taken to represent this lineation.

The variation in plunge of minor folds on the crest of the Enosburg Falls Anticlinorium becomes greater on its eastern limb (Fig. 6, domains 9b, 10b and 11b) and the folds become markedly disharmonic (Fig. 14), especially in the Mansville facies (Fig. 3 and 11f) (see below). In the Richford-Valcourt Syncline, steeper southerly-plunging folds predominate (Fig. 6, domains 12b and 13b); on the eastern limb, as dips flatten into the Sutton Anticline, the early folds change pitch to become reclined and trend more westerly (Fig. 6, domain 13b). The folds are strongly asymmetrical with attenuated limbs and thickened crests up to 2 m across (see Fig. 15h,i). Observations are limited by inadequate outcrop and the tightness of folds, but these minor folds appear to maintain a constant "Z" vergence across the core zone from the eastern to western limb (Fig. 5), a feature that indicates that they may be later than the main syncline, and at one outcrop a refolded fold was observed (Fig. 13c). These relationships are discussed later.

Brome Slide Zone

The attenuation of Oak Hill strata into the Mansville facies, mentioned previously, is to a large extent structurally controlled, inasmuch as the thinning is greatest on the eastern limbs of each synclinal outlier across the crest of the Enosburg Falls Anticlinorium (see cross section). This indicates that the thinning of the sequence is

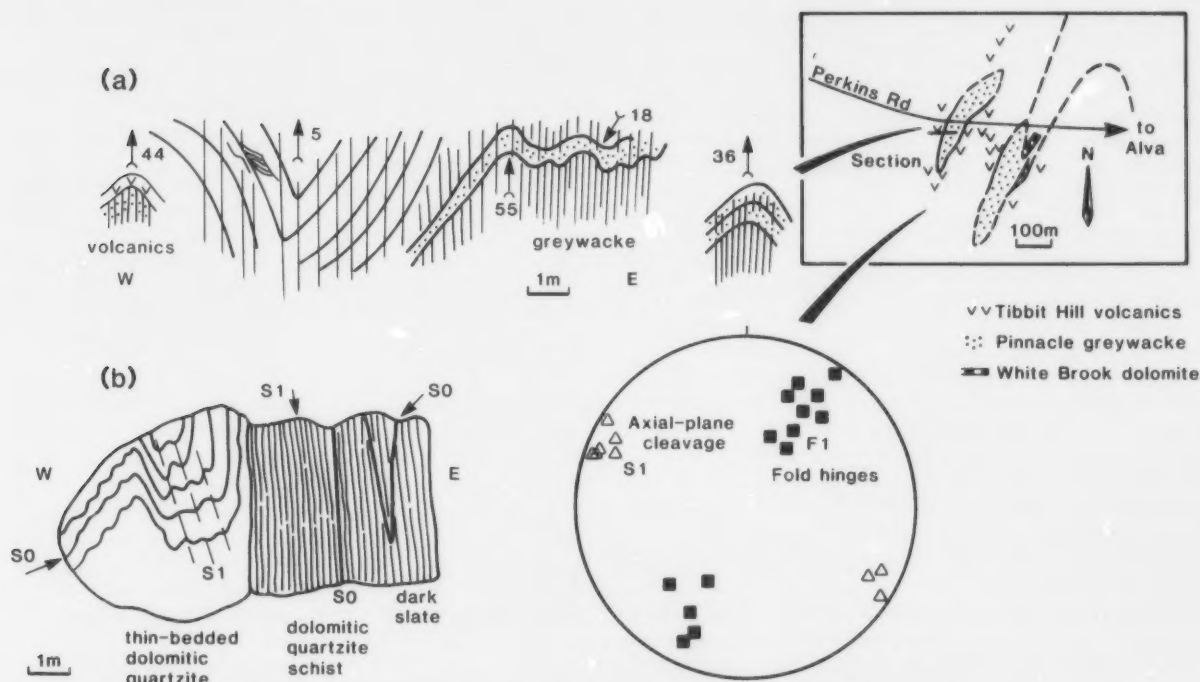


Figure 14. Plunge variations and disharmonic relationships in early folds. **a)** Map of tight infolds of Pinnacle greywacke in Tibbit Hill volcanics (inset). The plunge variation is indicated on the cross section and stereogram. Location—Perkins Road, 2.3 km west-northwest of Alva (830935). **b)** Dunham (?) dolomite, quartzite and slate disharmonically infolded into the Bonsecours schist. Location—0.5 km west of Abercorn (837891).

due largely to a layer-parallel attenuation during the early folding. Such a feature is reminiscent of the slide thinning on limbs of Dalradian recumbent folds (Rickard, 1962; Hutton, 1981). The structural style of the Mansville facies is intermediate between that of the Enosburg Falls Anticlinorium and the Richford-Valcourt Syncline. Greywacke and dolomite layers are boudinaged and sheared out in several places, and tight synclines are separated by even tighter sheared out anticlines. Fold-plunge variation is great (Fig. 3, 5 and 6, domain 10b) and steep folds are common especially in incompetent rocks. Folds are tight and commonly attenuated or sheared out along their limbs (Fig. 3 and 11e,f). The reappearance of units like the Tibbit Hill greenschist along the line of an anticlinal-hinge trace (Fig. 3a), and the long string of outcrops south of the main one at Mansville suggests that the overall structure is gently plunging.

Clark (1934) and Cooke (1954) postulated an underthrust (Brome Thrust) of the Sutton schists beneath the Oak Hill Group in this position, and Osberg (1965) mapped westward-dipping thrusts along this line to the north. To the south Stanley and Ratcliffe (1985) have mapped an eastward-dipping fault in this position, separating the Hinesburg and Underhill thrust slices. That there is a major lithological or structural break in this position is indicated by the steep gradient in the aeromagnetic contours on the Sutton aeromagnetic map (Geological Survey of Canada, 1954a). Nevertheless it is thought that the structure is a slide, or complex strain discontinuity (Fleuty, 1964b; Hutton, 1979), rather than a thrust fault. Attenuation of the eastern part of the Oak Hill Group prior to and during F_1 folding might be related to early westward overthrust movements in a similar manner to that described by Hintze (1978) from the Sevier Orogen in western Utah. The original slide zone later became modified by tight variably-plunging folds.

Sutton Anticline-early structures

The earliest structural element that can be observed in the Sutton schists is a micaceous schistosity (S_0) parallel to lithological layering. The structure is enhanced by segregations of quartz so that thin platy veinlets almost everywhere parallel the schistosity (Fig. 15a,d,e,g). Bedding schistosity has been reported elsewhere in the Green Mountains to the south in Vermont (e.g. Osberg, 1952; Chidester, 1953; Albee, 1957, 1968; Christman, 1959; Christman and Secor, 1961), and to the north in the Bennett schists (St. Julien and Hubert, 1975; St. Julien et al., 1983). In several of these examples, however, the schistosity contains quartz rods that correspond to the early east-west folds in the Sutton area so that it is not a true bedding schistosity. In the Sutton Mountains, although there are certain irregularities (described below), nowhere are there intrafolial folds with this schistosity in their axial planes, and the earliest axial-plane foliation is everywhere a crenulation cleavage. Moreover, several thin marker beds of greenschist in the bedding schistosity have been traced over many hundreds of kilometres and the map patterns show no sign of earlier major folds (Doll et al., 1961; Osberg, 1965).

The earliest folds (F_{1X}) observed are a series of small east-west recumbent and reclined folds that occur throughout the Sutton schists (Fig. 5). They plunge west on the western limb of the Sutton Anticline (Fig. 6, domains 14b to 18b and 20b to 21b), and cross the anticline (domain 19b) to plunge east on the eastern limb (domains 22b to 24b). Similarly, the associated gently-dipping axial-plane crenulation cleavage (S_{1X}) is folded across the anticline, and plots with the bedding schistosity (domains 14b to 24b). In the limbs of folds this cleavage is usually indistinguishable from the bedding schistosity. Its relationship to porphyroblasts and other textural details are described in the section on Metamorphism.

Folds are tight to isoclinal with strongly attenuated limbs (Fig. 15). Quartz veinlets are folded, lineated and in places shredded out into the foliation and almost transposed (Fig. 15d); they provide quartz-rod lineations (L_{1X}) parallel to the F_{1X} fold hinges (see Fig. 7e) and only rarely is the lineation slightly oblique to the hinge (see Fig. 18k). These folds are well exposed and particularly abundant in Glen Sutton and Bolton Pass where they form cascades several tens of metres across, as in Figure 15b; the triangular-shaped outcrop of the Mount Echo greenschist body (see Map 1750A) is a fold-hinge zone. The northward narrowing of the Sallys Pond formation at Roundtop and Dunkin (see Map 1750A) may be due in part to large east-west folds, and in the vicinity of Bolton Pass and Bolton Glen several steep dips at the nose of large early folds trend perpendicularly across the Sutton Anticline (see Map 1750A and Fig. 6, domains 14b, 19b and 22b).

Traverse mapping in the thickly timbered mountains did not allow the tracing of marker beds around any large folds except the Mount Echo fold; the possibility of detecting larger folds is frustrated by the inconstant vergence of minor folds, and in some places, even conjugate relationships were seen. Some examples of such irregular folds are illustrated in Figure 16, but they are exceptions rather than general features. Bedding schistosity and quartz veinlets pass round the hinges of all the folds, even those that show refolded patterns. These are best explained as local polyclinal developments in essentially the same phase of folding.

Nowhere in the Sutton schists was evidence found for early northerly-trending folds recumbent or overturned toward the west such as are reported in relation to early nappe folds in the Bennett schists to the north (St. Julien and Hubert, 1975; St. Julien et al., 1983). The serpentine lenses within the Sutton schists may mark the trace of early thrusts (Doolan et al., 1982; Stanley and Ratcliffe, 1985), but in the Sutton area there is no evidence of specific early structures related to these lenses.

Eastern Complex-early folds

The style of early minor structures in the Eastern Zone is illustrated in Figure 17. Although Lamothe (1981) mapped the contact as a fault, structurally there is no difference between the Bonsecours and Sweetsburg rocks (Fig. 17b,d) on the eastern limb of the Sutton Anticline,

and there is strong folding along the contact between the two formations. As the contact is approached, the axial-plane foliation steepens and the reclined folds become nearly vertical (Fig. 6, cf. domains 22b,c to 24b,c with 25b,c). In the less metamorphosed Sweetsburg Formation, the competent quartzite beds define steep tight asymmetrical folds (Fig. 17a,b) although they break up into

disconnected slabs and lenses (Fig. 17f); plunges vary from steep to moderate. Axial-plane foliation varies from a wide-spaced fracture cleavage through fine crenulation cleavage to a nearly slaty cleavage. Fold vergence is variable (Fig. 5) but "S" folds predominate, especially in the south. Very rarely a moderate north-plunging fold has a steeply crenulated axial foliation that may indicate that

Figure 15. Style of early cross folds (F_{1X}) in the Sutton schists. Bedding and bedding schistosity (S_0) are folded by F_{1X} , with S_{1X} as axial-plane crenulation cleavage. **a)** "Z" cross fold in quartz veinlets (black) parallel to bedding (S_0) in the Bonsecours schist. Late crenulation cleavage (S_2) is locally developed. Location—Alva (859939). **b)** Large east-west cross folds in thin quartzite layers in graphitic albite schists of the Mohawk Lake formation. Location—Benoit Road, 5 km east of Sutton Junction (940018). **c)** Cross folds in quartzite (stippled) and albite gneiss beds of the Sallys Pond formation. Note the alignment of albite porphyroblasts (black) across the S_{1X} foliation. Location—Highway 105A, East Richford (903874). **d)** Cross folds in quartzite (stippled) and shredded quartz veinlets (black). Location—Roundtop Mountain (931946). **e)** Shredded quartzite layers (stippled) and quartz veinlets (black) in Mohawk Lake formation. Location—Eberdt Road, 4.5 km east-northeast of Sutton (920984). **f)** Tightly folded quartzite schist layers (stippled) of the Mohawk Lake formation. Location—1.7 km east-southeast of Sutton (892968). **g)** Folded schistose quartzite layers (stippled) disrupted along S_{1X} crenulation cleavage, Mohawk Lake formation. Location—3 km west-northwest of Glen Sutton (905912). **h)** Intrafolial "Z" cross folds in steeply-dipping quartzite schist layer, Bonsecours schist. Location—2 km south of Sutton (877953). **i)** Tight "Z" cross fold in Bonsecours schist. Location—2 km south of Sutton (877953).

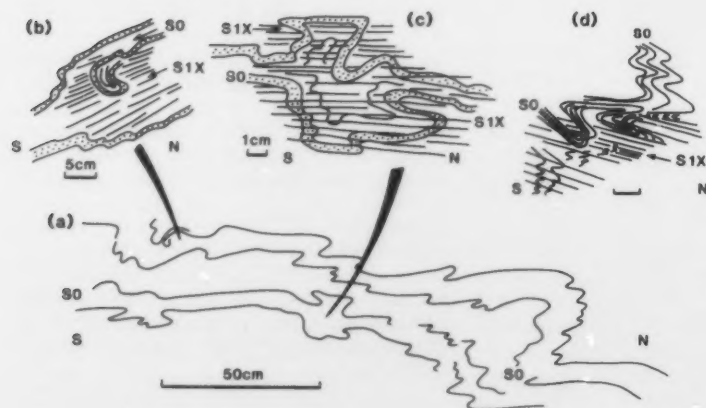
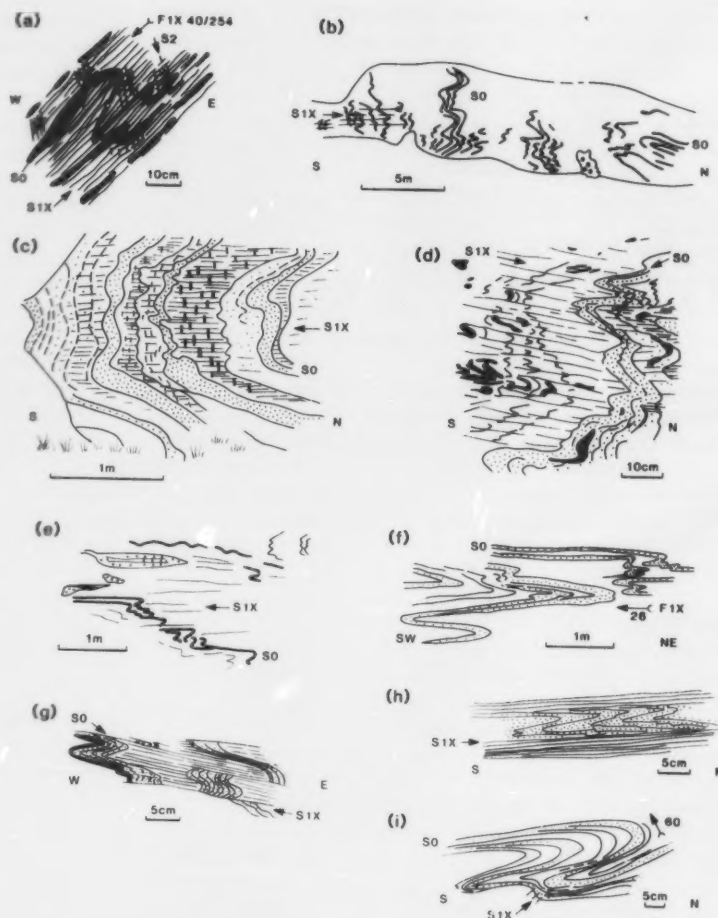


Figure 16. Incongruous relationships in early cross folds. **a)** Folded quartzite layers (stippled) in albite gneiss of Sallys Pond formation. Location—Highway 105A, East Richford (902873). **b)** and **c)** Detail from **a)**, showing the incongruous crosscutting relations of S_{1X} crenulation cleavage. **d)** Crosscutting relations of S_{1X} crenulation cleavage. Location—Highway 105A, Glen Sutton (93405).

the steep folds were superimposed on earlier less steep ones as in the Richford-Valcourt Syncline, but insufficient examples were found to establish the relationship satisfactorily. No extension lineations were recognized in the Eastern Zone and no evidence was found for rotation of fold axes; the folds are thus most likely equivalent to the cross folds (F_{1X}) of the Sutton Anticline.

The serpentinite belt marks a zone of strong tectonic disturbance according to Laurent (1975) but exposures are inadequate in this area to make any useful deductions. An important observation, however, is that thin layers of serpentinite are folded and brecciated by the steep folds (Fig. 17j). Moreover, the Baker talc-serpentinite lens (2 km west of Bolton-South) and other small lenses along the eastern limb of the Sutton Anticline appear to be located in the hinges of early steep folds, and their margins are folded. These observations confirm the relationships described by St. Julien (1967) but conflict with those of Chidester (1968) who argued that the ultramafics in Vermont were intruded after early folding.

Major steep reclinined folds in the St. Daniel Formation are defined by the Bolton pillow lavas. Fold hinges at Sugarloaf and Pevee mountains and 3 km west of Vale Perkins constitute a large complex fold structure (see map). West of Vale Perkins, the vergence appears to

reverse; Lamothe (1981) mapped a fault offset, but in this report the structure is interpreted to be a giant "conjugate" fold similar to those in the Sutton schists (Fig. 7g and 16). Observations on pillow lavas indicate that this belt youngs consistently eastward. It is not possible to determine how this northern belt of volcanic rocks links with the masses to the south, and aeromagnetic anomalies on the Memphremagog aeromagnetic map (Geological Survey of Canada, 1954b) do not support joining the lavas across the Vale Perkins Fault as was done by Ambrose (1942). The mass at Bear Mountain appears to young northward and there is a steep fold closure south of the International Boundary (Cady et al., 1963). The belt south of Owl Head Mountain continues for 8 km south into Vermont (Doll et al., 1961), but its relationship to the Bear Mountain synform is unclear; pillow lavas indicate a westward younging (see map) and minor folds on the eastern side show a steep southerly plunge and S-vergence (Fig. 5 and 17i). The Bear Mountain fold cannot be equivalent to the late Sargent Bay Syncline because of its steep plunge (Cady et al., 1963), but further work is needed to determine whether or not the Bear Mountain mass is an offset continuation of the Sugarloaf-Pevee layer, and whether it has been folded around the Sargent Bay Syncline; as explained later, refolding is considered unlikely.

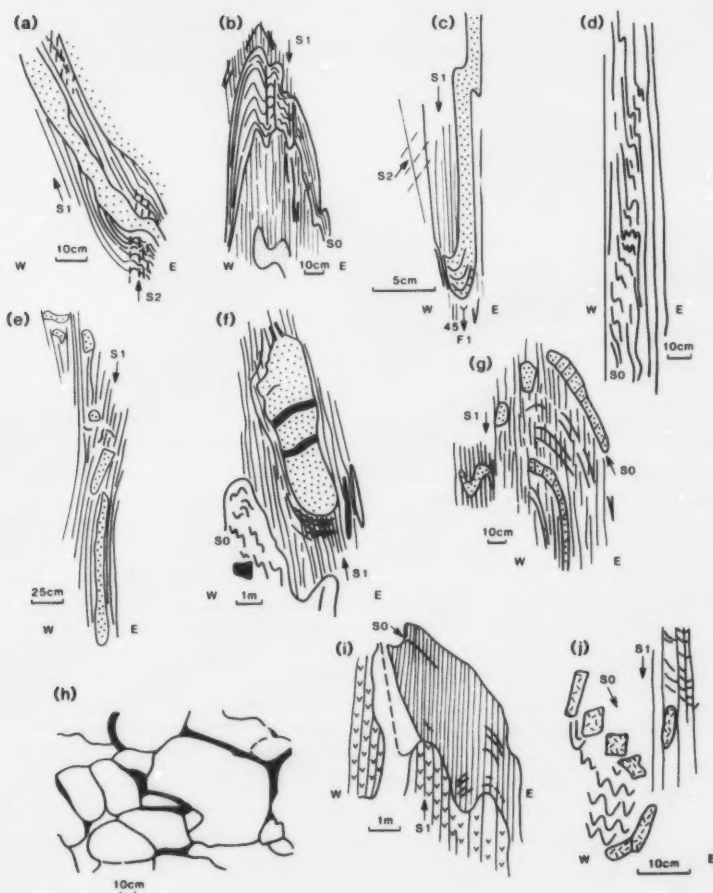


Figure 17. Style of early minor structures in the Eastern (Mansonville) Zone. **a)** Quartzite beds (dotted) in Sweetsburg slates with early slaty cleavage (S_1 or S_{1X}) and late crenulation cleavage (S_2). Location—Highway 243, Mansonville (052912). **b)** Steeply-plunging fold and slaty cleavage (S_1) in quartzite and slate of the Sweetsburg Formation. Location—3 km north of Mansonville (049947). **c)** Disrupted fold in grey slate of the St. Daniel Formation. Note the weak trace of late crenulation cleavage (S_2). Location—1.5 km south of Potton Springs (076006). **d)** Steep intrafolial "S" folds in quartzitic schists of the Bonsecours Formation. Location—4.5 km north-northwest of Mansonville (045960). **e)** Shredded quartzite beds in the St. Daniel slates. Location—2.4 km east of Sugarloaf Mountain (118995). **f)** Folded and disrupted quartzite (dotted) with quartz veins (black) in the Sweetsburg slate. Location—5 km north of Mansonville (048964). **g)** Steep folds in disrupted quartzite (dotted) in black slate of the St. Daniel Formation. Location—Sugarloaf (099992). **h)** Pillow lavas in Bolton volcanics. Location—lakeside, 1.5 km southeast of Vale Perkins (135955). **i)** Steep folds at the contact between the Bolton volcanics (V pattern) and the St. Daniel slate. Note the cleavage in the volcanic rocks and the weak trace of bedding in the slates. The contact at this locality was claimed to be unconformable by Cooke (1950), but is disputed here. Location—lakeside, 1.5 km southeast of Vale Perkins (135955). **j)** Disrupted serpentinite fragments (dash pattern) in steeply-plunging early fold. Location—Peabody (067947).

Minor structures in the St. Daniel slates are all moderately to steeply plunging (Fig. 17c,g) and exactly match the pattern of those in the Sweetsburg (Fig. 17b); they are plotted together on Figure 6, domain 25. The poles to bedding clearly define a steeply-plunging fold axis. Minor folds are common especially in the cores of the Pevee and Sugarloaf folds (Fig. 17g). A shaly foliation parallels bedding where thin quartzite interbeds exist and the axial-plane structures vary from none through a wide-spaced fracture cleavage to a good slaty cleavage. Fold hinges are often shredded to form pseudo-breccias (Fig. 17e,g). The massive volcanic rocks are roughly schistose at their margins and the adjacent shales are disharmonically folded and strongly cleaved (Fig. 17i).

Late structures

Two major structures developed later than those previously described—the Sutton Anticline and the Sargent Bay Syncline (Fig. 1 and 5). The former is the dominant structure in the central part of the area and its effect spreads to peripheral areas, and the latter affects Silurian rocks in the vicinity of Lake Memphremagog; there is no structural relationship between the two folds but they may be of the same age.

Sutton Anticline

It is proposed to use the term Sutton Anticline for the specific late structure in this area and Green Mountain-Sutton Mountain Anticlinorium (Cady, 1969; Osberg, 1969) for the general regional structure. It is convenient to describe the main anticline first then to consider related structures in the Enosburg Falls Anticlinorium and the Eastern Complex.

The anticline is well defined by an arching of the bedding-schistosity (S_0) and axial-plane foliation (S_{1X}) to early folds (Fig. 5 and 6, domains 14b,c to 24b,c). Associated minor folds of chevron style (Fig. 18a,b) occur in limited zones across the anticline, and these clearly re-fold the pre-existing planar elements (S_0 , S_{1X}) and quartz-rod lineations (L_{1X}) (Fig. 7d,e,f and 18). The folds are accompanied by an axial-plane foliation that varies from a spaced fracture cleavage (Fig. 18b,c,d) to a crenulation cleavage (Fig. 18e and 30a) (Rickard, 1961, 1965).

In zones where the bedding schistosity is steep or large early folds predominate, the late folds are represented by crenulations. Because of the high-angle relationship between the late axial plane (S_2) and early folds (F_1), some late crenulations plunge steeply as they cross the early fold hinges (Fig. 7f and 18k); however, few of these were measured.

The vergence of the late minor folds shows them to be congruent with the major anticline, and the axial-plane foliation dips steeply to the east-southeast on both limbs as the fold is slightly asymmetrical (see cross section and Fig. 6, domains 14a to 24a). This foliation (S_2) is not exactly parallel to the trace of the hinge plane of the major fold, however, so that the minor folds plunge northward

on the western limb and southward on the eastern limb (Fig. 5 and 6, cf. domains 14b,c to 18b,c with 22b,c to 24b,c). This transecting arrangement probably results from the late development of the cleavage in the folding process (Fig. 19). A similar arrangement has been described in New Brunswick (Stringer, 1975; Borradaile, 1978).

Influence of the Sutton Anticline on adjacent areas

A late crenulation cleavage or spaced fracture cleavage cuts the early foliations in the Richford-Valcourt Syncline and the Enosburg Falls Anticlinorium. It cuts across both limbs of the steep folds in the syncline and the variably-plunging folds in the anticlinorium (Fig. 12b,g,h, 15a and 18i,j). Late fold hinges generally have a gentle plunge except where they cut across steep early folds (Fig. 7a,c and 6, domains 5c to 13c). Only definite early lineations were plotted as such in Figures 5 and 6, so that some of the steeper lineations classified as late in these domains may indeed be early ones. In places in the Sutton valley, the late crenulation cleavage in the Bonsecours schist becomes as strong as the early crenulation cleavage (Fig. 18e). Late cleavage is locally strong where stress perturbations developed around vein-quartz knots or early fold hinges (Fig. 12c, 15a and 33c). In most of the rocks of the Oak Hill Group the late cleavage is a weak fracture cleavage or is represented only by crenulations. In some schistose greywacke layers in the eastern outcrops, a "heringbone" effect is created by the differential development of early and late foliations in adjacent layers of different composition (Fig. 18h). The late foliation is represented by weak kink planes in the lower Gilman phyllites and not at all in the upper Gilman quartzites; it is more intense in the Sweetsburg Formation (see Fig. 34) especially where the gently-dipping early cleavage is buckled against steep beds in the noses of the Oak Hill and Hunter Mills anticlines (Fig. 9d). Late folds are well developed in the Stanbridge slates and, in the Rioux Quarry (Williams, 1966), the crenulation cleavage almost obliterates the early slaty cleavage and crenulates the crosscutting quartz veinlets (Fig. 8e and 9). A second crenulation cleavage locally crosses the first and the two are thought to have developed sequentially during progressive deformation (Helmstaedt and Dixon, 1980).

The cleavage profile on the geological cross section on the map shows the relationship of the late crenulation cleavage to the early foliation. The crenulation cleavage is vertical or dips steeply to the east across the Richford-Valcourt Syncline and the Enosburg Falls Anticlinorium and it strikes consistently slightly to the north of the early foliations; this feature is illustrated by transferring the maxima points of the late cleavage poles to the early cleavage plots in each domain (Fig. 6, domains 5a,c to 13a,c). Minor folds associated with the late crenulation cleavage show a vergence sense consistent with the western limb of the Sutton Anticline across both limbs of the Richford-Valcourt Syncline and the crest of the Enosburg Falls Anticlinorium. The same relationship is reported by Marquis (1985) from the Richmond area to the north. Where the early cleavage flattens

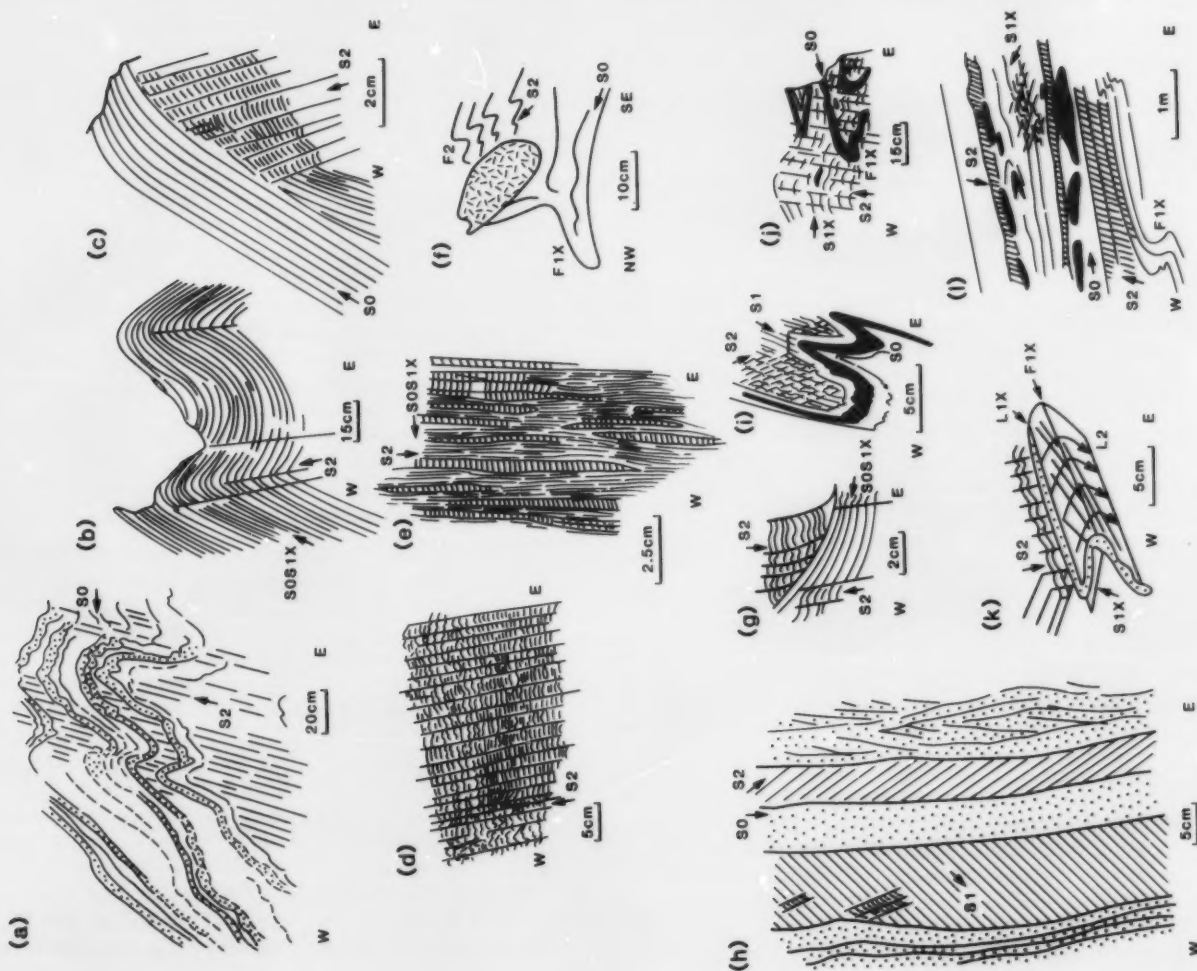


Figure 18. Style of late structures and superposed relationships relating to the development of the Sutton Anticline. Late folds (F_2) and crenulation cleavages (S_2) are superposed on pre-existing bedding schistosity (S_0) and the early axial-plane structures (S_1 and S_{1X}). a) Folds and crenulation cleavage (S_2) in Sallys Pond gneiss. Location—Bolton Pass, Highway 243 (021045). b) Chevron folds in Mohawk Lake graphite schists. Location—intersection of Eberdt and Kelly roads, 3.5 km east-northeast of Sutton (911981). c) Differential development of bedding schistosity (S_0) and late crenulation cleavage (S_2) in Bonsecours schist. Location—2 km southwest of Sutton (860962). d) Fracture style late crenulation cleavage (S_2) in Mohawk Lake schists. Location—3.5 km east of Abercorn (878895). e) Small-scale transposition cleavage with quartz and mica-rich layers. Parallel to the late crenulation cleavage (S_2) in Bonsecours schists. S_0 - S_{1X} cross foliation developed in quartz-rich layers. Location—2 km southwest of Sutton (860962). f) Albite-rich segregation knot (dashed) in F_2 fold in Sallys Pond gneiss. Location—3 km southeast of Glenton (951874). g) Differential development of S_2 crenulation cleavage and early foliation (S_0 , S_{1X}) in Bonsecours schist. Location—1 km west of Sutton Junction (878020). h) Differential development of slaty cleavage (S_1) and crenulation cleavage (S_2) in different beds (S_0) of Pinnacle grey-wacke; quartzite beds (stippled), siltstone with S_1 and mudstone with S_2 . Location—Perkins Road, 4.5 km north-northwest of Abercorn (829936). i) Late crenulation cleavage (S_2) crossing the axial-plane foliation (S_1) of an early fold in laminated chlorite schists of the Tib-bit Hill volcanics. Location—Mansville (877026). j) Early cross folds (F_{1X}) with axial-plane cleavage S_{1X} refolded by late folds with axial-plane crenulation cleavage (S_2) in Bonsecours schist. Location—2 km southwest of Sutton (860963). k) Late crenulation cleavage (S_2) and lineations (L_2) cutting early cross folds (F_{1X}) in Mohawk Lake schists. Note the discordant L_{1X} lineation. Location—5 km east of Sutton Junction (942021). l) Late crenulation cleavage (S_2) cutting across early cross folds (F_{1X}) in Bonsecours schist. Quartz veinlets (black). Location—3 km northeast of Sutton Junction (910039).

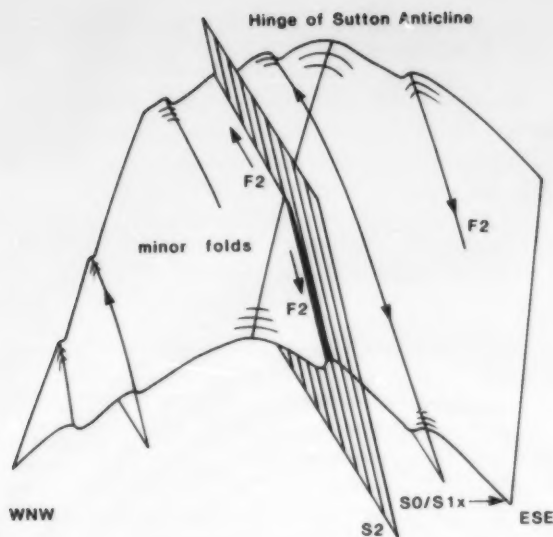


Figure 19. Diagrammatic representation of the transecting nature of the late crenulation cleavage (S_2) in the Sutton Anticline. Note that the minor F_2 folds plunge to the north in the western limb and south in the eastern limb (cf. Fig. 5 and 6).

and then dips eastward, the vergence of minor crenulations changes as if a broad late synform existed to the west of the crest of the Enosburg Falls Anticlinorium. This is not a true fold, however, because the fanning of the early cleavage was not caused by the late structure. The relationships described above clearly establish that the final form of the Sutton Anticline and its related minor structures were developed later than the Richford-Valcourt Syncline and the Enosburg Falls Anticlinorium. The late deformation may have tightened the early folds somewhat, but the three major structures cannot be directly related as concluded by Eakins (1964) and Osberg (1965).

On the eastern limb of the Sutton Anticline the late crenulation cleavage weakens and dies out; it affects the Sweetsburg slates slightly and although it is definitely developed east of the serpentinite belt its occurrence there is rare, and it was nowhere seen east of the outcrop of the Bolton Member (Fig. 5 and 6, domain 25). There is a weak crenulation in places in the St. Daniel slates, but this appears to be related to a northeast-trending cleavage that also affects the Silurian rocks (Fig. 6, domains 25 and 26) (see also Lamothe, 1981).

Sargent Bay Syncline

The Sargent Bay Syncline is the westernmost of a series of folds in Silurian and Devonian rocks of the Connecticut Valley-Gaspé Synclinorium (Fig. 1). The Sargent Bay fold is a simple upright syncline and its hinge is well exposed 1 km southwest of Vale Perkins. There is an axial-plane slaty cleavage, and bedding planes trace across it to give a prominent lineation that indicates local moderate plunge variation from 0 to 30 degrees north (Fig. 6,

domain 26). This contrasts markedly with the reclined folds in the St. Daniel Formation beneath the unconformity (Fig. 5 and 7g). It is important to note that the cleavage in the Silurian slates is parallel to that in the St. Daniel (Fig. 6, cf. domains 25 and 26) but there is generally no second cleavage below the unconformity. The unconformity cuts across a thin layer of the Bolton volcanics above the main Sugarloaf layer and the crosscutting nature of the contact is indicated by the different units below the Peasley Pond conglomerate or Glenbrook slate.

Faults and joints

Apart from the early faults (Hinesburg Thrust, Brome Slide Zone and the serpentinite zones) previously described, faulting appears to play a minor role in the deformation of the rocks in this area. The few small cross faults (see Map 1750A) are nowhere exposed and they were mapped on the basis of offsets of beds. The Vale Perkins Fault in the southeastern corner of the area is more hypothetical, although it is marked by sharp truncations of magnetic contours on the Memphremagog aeromagnetic map (Geological Survey of Canada, 1954b). It may be a large wrench (strike-slip) fault separating disparate portions of the Bolton volcanics; it does not offset the serpentinite belt, however. A parallel structure passes through Mountain House Wharf where Devonian rocks are wedged between Bolton volcanics.

Joints were analyzed from the two zones mapped in greatest detail. In the Enosburg Falls Anticlinorium the joint pattern is asymmetrical (Fig. 20a). A major northwest-trending set (a) is oblique to orthogonal sets (b and c) of east- and north-trending joints. The east-trending set (b) possibly comprises small-angle conjugate sets, but neither of these is exactly perpendicular to the general fold-axis trend (Fig. 20a). Charbonneau (1975b) has recorded similar north- and east-trending orthogonal joint sets from the Frelighsburg area to the west; however, the northwest-trending set is not developed there.

The joint pattern from the Richford-Valcourt Syncline and the western limb of the Sutton Anticline (Fig. 20b) is similar to that from the Enosburg Falls Anticlinorium except that the north-northwest-trending oblique set (d) is weaker and the north-northeast-trending set (a) is stronger. The latter is parallel to the late crenulation cleavage. The separation of approximately east-trending sets (b and c) is better defined here, and the west-northwest-trending set (b) is perpendicular to the axis of the late fold system. The east-northeast-trending set (c) and the north-northwest-trending set (d) may represent a conjugate system about the west-northwest-trending joints (b), but if so they are not coaxial with the orthogonal sets (a and b). It is not clear how these joints relate to either of the two systems of folds.

Quartz veins are developed in some joints of all of the sets, but they are most common in the north-northeast-trending set that is parallel to the late (S_2) cleavage.

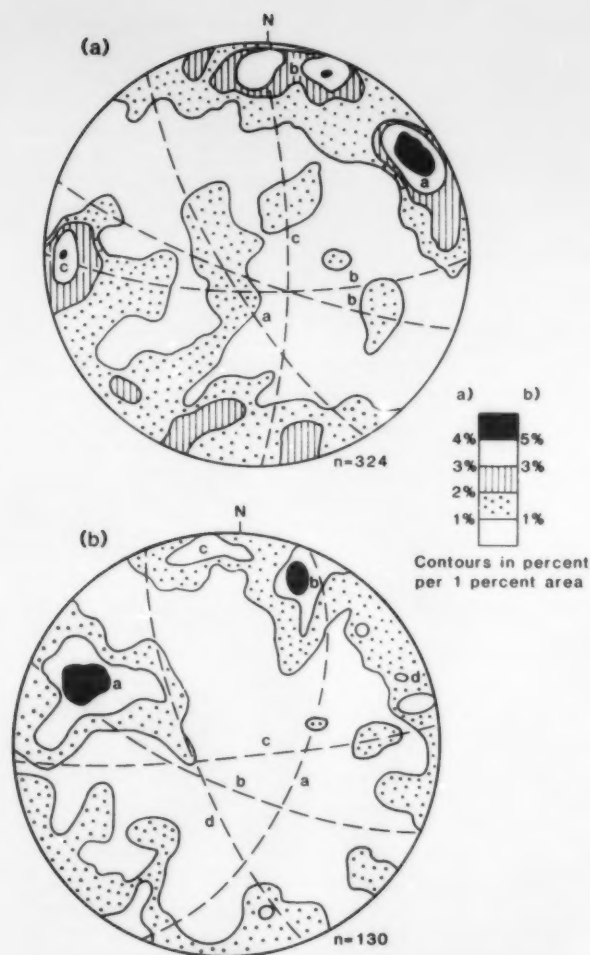


Figure 20. Stereograms of joints plotted on the lower hemisphere of a Schmidt equal-area net. **a)** In the Enosburg Falls Anticlinorium. **b)** In the Richford-Valcourt Syncline and the western limb of the Sutton Anticline. Great circle planes labelled a, b, c and d are mean joints corresponding to maxima on pole plots.

Geometry and correlation of folds, and mechanism of formation

The outstanding feature of the fold geometry of this area is the contrast between east-west reclined cross folds in the Sutton Anticline and the northerly-trending but variably-plunging folds in the Enosburg Falls Anticlinorium. These two phenomena are probably unrelated but a major problem is the extent to which the cross folds penetrate outward into the steeply-dipping eastern and western flanks of the Sutton Anticline.

The correlation of folds on the basis of geometry (style and orientation) is fraught with difficulty (Park, 1969; Williams, 1985), especially in cross-fold situations. Nevertheless, the fact that two phases of folding can be recognized in each of the four structural zones, and that the latest structures have the same style and orientation

and they appear to be related congruously to a single major structure, suggests the possibility that the early structures in each zone may also be correlated in spite of the considerable differences in their geometry and trend (Table 3).

The Enosburg Falls Anticlinorium essentially comprises an array of northeast-trending, open to tight, gently-plunging folds with upright axial planes, which in the west become overturned to the west (see cross section). In detail, many minor folds, and even larger ones controlling lithological boundaries, have a moderate to steep plunge (Fig. 6, domains 4b to 13b). The steep stretching lineations (L_{1a}) in the foliation planes show little variation (Fig. 6, domains 4c to 13c). The association of steeper plunges with tighter folds in the less competent rocks suggests that the variation in fold plunge resulted from a flattening process with rotation of hinges toward the "stretching" direction (Sanderson, 1973).

The extreme variations and preponderance of very tight steep folds in the Mansville facies (Fig. 6, domain 11) coupled with the evidence for structural attenuation indicates that there is a zone of marked strain discontinuity (Brome Slide Zone) along the eastern side of the Enosburg Falls Anticlinorium. The rotation of fold hinges is particularly effective in shear zones (Bryant and Reed, 1969; Garnett and Brown, 1973; Escher and Watterson, 1974; Cobbold and Quinquis, 1980). To the east of this zone, in the Richford-Valcourt Syncline, most of the minor folds are steeply plunging in a near-vertical axial plane (Fig. 6, domains 10b, 12b and 13b). Although the strain discontinuity appears to act as a transition zone, the correlation of these steep folds with the variably-plunging but essentially north-trending folds (Fig. 6, domains 5b to 8b and Fig. 7b,c) in the Enosburg Falls Anticlinorium is difficult. Many of the steep folds almost certainly are the same as the east- and west-trending reclined folds (F_{1X}) that cross the Sutton Anticline (Fig. 7c,f); they have the same style and orientation.

The consistent "Z" vergence of the steep minor folds across the Richford-Valcourt Syncline (Fig. 5) indicates that the folds are superimposed structures (Fig. 21). Support for this interpretation comes from the few examples along the eastern side of the Enosburg Falls Anticlinorium where the earliest foliation is folded about steep axes (Fig. 13) and from the previously mentioned stratigraphic

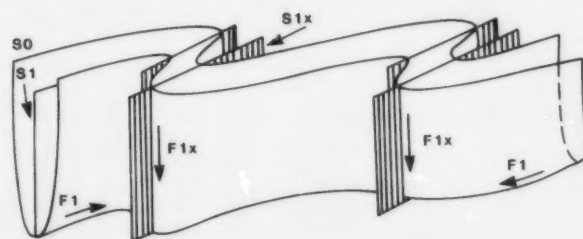


Figure 21. Superposed minor fold developments in the Richford-Valcourt Syncline. Superposition of the S_{1X} axial plane across the F_1 fold limbs give steep F_{1X} minor folds with a constant vergence across the major fold.

evidence that the Richford-Valcourt Syncline is essentially a gently-plunging structure. The latter point is strengthened by the fact that to the north of this area the Richford-Valcourt folds flatten to a northwest-dipping axial surface and are themselves folded across the Sutton Anticline (Osberg, 1965; Marquis, 1985), where they have a different trend from the cross folds (Fig. 22a).

The steep folds on the eastern flank of the Sutton Anticline have a similar relationship to, and are probably also younger than, the gently-plunging north-trending folds. Lamothe (1981) correlated the steep reclined Sugar Loaf and Pevee Mountain folds with the late structures, but these folds, although larger, are geometrically similar to the reclined folds in the Sutton schists and different from the late folds. In the Bonsecours schist, steep folds (F_{IX}) and a crenulation cleavage (S_{IX}) cut the bedding schistosity and are themselves crenulated by the late cleavage (S_2), as on the western flank of the Sutton Anticline.

Thus it is concluded that the reclined cross folds are younger than the north-trending upright folds. This also means that the development of bedding schistosity in the Sutton schists might be an equivalent phase (not necessarily contemporaneous) to the north-trending F_1 folds in the Oak Hill Group higher in the sedimentary pile (Table 3). The fact that quartz veinlets in the schists are parallel to the bedding schistosity whereas those in the upper part of the Oak Hill sequence are concentrated in or close to the S_1 axial-plane foliation (Fig. 8e,f) lends support for this contention.

Cross folds are a common feature of orogenic belts and they may develop sequentially by an interplay of complex movements, or contemporaneously (King and Rast, 1955; Rast and Platt, 1957; Brace, 1958; Rickard, 1963; Fyson, 1964; Haller, 1970). Cross folding occurs in the schists along the entire Green Mountain-Sutton Mountain Anticlinorium (Brace, 1953; Christman, 1959; Christman and Secor, 1961; de Römer, 1961; Osberg, 1965; Stanley and Ratcliffe, 1985). Chidester (1953) was probably the first to determine that these folds are older than the minor structures associated with the Green Mountain Anticline. They may be caused by variations in fold plunge along the main trend (Chidester, 1953), by movements of cover over an irregular basement (Brace, 1953, 1958) or by shear movements associated with the development of salients (Cady, 1969; Thomas, 1983). Unlike many cross-fold situations, these do not swing into the regional trend and they are everywhere crosscut by regionally trending later folds.

In the report area and in the Richmond area to the north (Osberg, 1965) there is a tendency for the F_{IX} folds to indicate northward overriding movements along the Sutton Anticline. There may be as yet unrecognized, larger early folds, but at the moment only small folds (a few tens of metres) are known and these commonly show inconsistent vergence (Fig. 5) and the tendency to "conjugate" geometry (Fig. 16) as also described by Brace (1953, 1958) from rocks close to the Grenvillian basement. A similar fold geometry may result from constraints deep

in a pile of sediments that are being overridden by an obducting sheet, especially if this was accompanied by bending movements associated with the development of salients and re-entrants.

Away from the Sutton Anticline proper, related late folds occur in several places. In the Stanbridge slates at Cowansville, late folds become very strong and almost obliterate earlier structures. Nevertheless one can identify three phases (Helmstaedt and Dixon, 1980); whether or not the early F_1 phase in these Stanbridge rocks is equivalent to those in the Sweetsburg (Fig. 9) depends on the existence and intensity of the supposed fault or disconformity between them. The quartz veinlets in or close to the early axial-plane foliation indicate a reasonable correlation for the late F_2 folds which crenulate the quartz in both formations, and suggest also a correlation of the F_1 folds. To the northeast in the Beranger area Charbonneau (1975a) reported three sets of folds in the Stanbridge rocks in addition to sedimentary folds. He mentioned that F_2 folds deform quartz veins so that these may correlate with F_2 in this area. An additional phase of late folding (F_3) is reported at Cowansville (Helmstaedt and Dixon, 1980) and is also observed in the Stanbridge and Granby nappes (St. Julien and Hubert, 1975).

To the northeast in the Bennett schists and Caldwell Group, St. Julien (1967) and St. Julien et al. (1972) reported two phases of folding before the late anticline, but the Caldwell in the Mansonville area shows only one phase (reclined) of early folding. Similarly Robinson and Fyson (1976) reported three fold sets in the Cambro-Ordovician Ascot Formation to the east. It is likely that a different sequence of events occurred in these areas, and St. Julien and Hubert (1975) have postulated that locally nappes slid southeastward.

The Magog Group strata (including the Beauceville Formation) lie unconformably above the St. Daniel Formation and were folded only once according to St. Julien et al. (1972); Robinson and Fyson (1976) on the other hand reported two phases of folding, one or both of which were thought to be later Devonian structures. Until this is settled it is difficult to attempt regional correlations. It is likely that the three phases reported here— F_1 , F_{IX} and F_2 —are equivalent to the three phases in the Bennett schists, but there is no basis for correlation at present.

The geometrical relations between the structures in the Cambro-Ordovician and Devonian rocks are enigmatic throughout the Canadian Appalachians (Rickard, 1965; St. Julien, 1967; Cady, 1969; Sikander and Fyson, 1969; Helmstaedt, 1970; de Römer, 1976; Robinson and Fyson, 1976; Hatch, 1982). The problem is well illustrated by the relations west of Lake Memphremagog where the gently-plunging Sargent Bay Syncline cuts across steeply-plunging folds in the St. Daniel Formation, although both sets of folds have parallel axial-plane slaty cleavages (Fig. 7g, S_{IX} and S_{ID}). The crosscutting nature of the unconformity (see map and St. Julien, 1965; Lamothe, 1981) precludes an interpretation by the contemporaneous development of main and cross folds, and the slaty cleavage (S_{IX}) in the steep tight reclined folds is most likely

older than that in the Sargent Bay Syncline (S_{1D}) (Fig. 7g). To the west, this early foliation is cut and folded by the crenulation cleavage of the Sutton Anticline which may be of Devonian age (Lamothe, 1981), since the metamorphism associated with the deformation has been dated (K-Ar) at 411-367 Ma (this report). Unfortunately the crenulation cleavage dies out eastward before the Sargent Bay Syncline so that it is impossible to determine the relationship between the crenulation cleavage of the Sutton Anticline and the slaty cleavage of the Sargent Bay Syncline.

The development of parallel fold systems of different ages is a common feature of orogenic belts and there is normally a basement-cover interaction with refolding or decollement (Ross and McGlynn, 1963). In the Memphremagog area there is no sign of refolding in the "basement" slates although the Devonian folding must have been stronger than a mild warping (Zen, 1972; Robinson and Fyson, 1976) to have developed a slaty cleavage in the Glenbrooke Group. A similar situation has been reported for the Devonian slates of the Portage Lakes area of New Brunswick (Helmstaedt, 1970). It is possible that the massive Bolton volcanics resisted further deformation but there is no sign that disharmonic folding occurred at the contact with the overlying Peasley Pond conglomerate to indicate decollement. The most plausible explanation seems to be that localized basement shearing along pre-existing cleavage planes allowed a syncline to develop in the Silurian cover rocks while the basement folds merely tightened. The unconformity stands as a monument to caution against correlating fold sets on the basis of similar trend.

In conclusion it is noted that the axial-plane trace or foliation of each successive deformation phase trends more northerly than the earlier one (Fig. 22a). The Enosburg Falls Anticlinorium and associated structures (S_1) trend slightly more northerly than the northeast-trending facies boundary on the Vermont-Quebec Geanticline

(Cady, 1969; Osberg, 1969) and the foliation (S_{1X}) associated with the steep folds in the Richford-Valcourt Syncline trend north-northeast across the earlier foliation (Fig. 22b). Then the Sutton Anticline refolds the Richford-Valcourt Syncline, and finally the latest foliation (S_2) transects the Sutton Anticline itself with a more northerly trend (Fig. 22b). Christman (1959) attributed a similar variation in Vermont to changes in the direction of compression. This slight but consistent variation might be better explained by postulating that the deforming welt was progressively bent as it "collapsed" against the re-entrant in the Precambrian Shield (Rankin, 1976; Doolan et al., 1982; Thomas, 1983). At each strain threshold, axial-plane foliations developed with the same north-trending orientation and were then progressively rotated clockwise in the northern limb of the Quebec re-entrant (Fig. 22b). Cady (1969) has attempted to explain the vergence of the steeply-plunging folds by a similar mechanism and Sanderson et al. (1985) have invoked rotation and strike-slip motions to explain transecting cleavage. Phillips et al. (1979) reported a similar transection caused by transpression along the Iapetus suture in Ireland.

METAMORPHISM

In this section of the report the metamorphic mineral assemblages are briefly described, and the relationship of mineral growth and quartz fabric to deformational features are examined. These relationships are summarized in Table 3.

Studies of the metamorphic assemblages in the Appalachians of northeastern United States have indicated the presence of two northerly-trending zones: a medium-pressure-series zone along the Green Mountain Anticlinorium and a low-pressure-series zone in eastern Vermont and New Hampshire (Thompson and Norton, 1968). The distinction is important because the metamorphism in the two zones is probably of different ages (Albee, 1968; Laird et al., 1984). The discoveries of

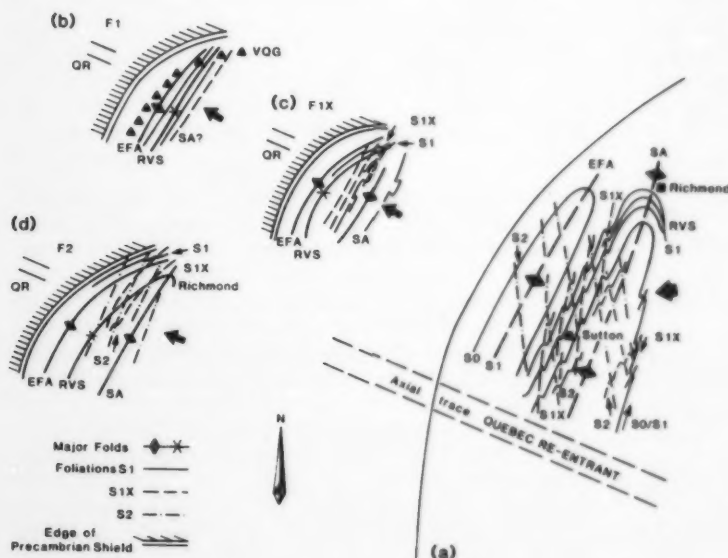


Figure 22. Plan of fold developments in relation to the Quebec re-entrant (QR). a) Present day areal relations of structures (angles between the traces of hinge surfaces are exaggerated). b), c) and d) Stages in the development of major structures as the orogenic welt is forced westward into the re-entrant in the Precambrian Shield. The Vermont-Quebec Geanticline (VQG, triangles) is crossed by F_1 folds (b) and F_{1X} cross folds (c) low in the sequence, and then crossed by late F_2 folds (d). EFA = Enosburg Falls Anticlinorium. RVS = Richford-Valcourt Syncline. SA = Sutton Anticline.

Table 3. Correlation of deformational and metamorphic events. Period boundaries based on Harland et al. (1982).

* Dated samples are listed in the Appendix.

These events may be Silurian with dates reset by a Middle Devonian reheating.

AGE	Ensbury Falls Anticlinorium	Mamville Zone	Richford-Vacourt Syncline	Sutton Anticline	Eastern Complex	Sargent Bay Syncline
Middle Devonian 380 Ma	Joints Cross faults	Joints Cross faults	Joints	Joints and quartz veins	Joints Cross faults CRENULATIONS Weak crenulation cleavage dips to southeast, east side zone only	Joints Cross faults CRENULATIONS Weak crenulation cleavage dips to southeast
Devonian 410 Ma	Minor quartz veins Local recrystallization chlorite and tourmaline; * (411-382 Ma)	[M ₂]#	Recrystallization — weak; micas in (S ₂) crenulation cleavage, small albite porphyroblasts, local biotite.	Recrystallization — strong; albite, micas, tourmaline, garnet, chlorite porphyroblasts * (380-387 Ma)	[F ₂]#	[F ₂]#
?	MINOR NORTH-SOUTH FOLDS upright chevron style fracture to crenulation cleavage	[F ₂]# [S ₂]	MINOR NORTH-SOUTH FOLDS consequent "Z" vergence; crosscutting crenulation cleavage (S ₂)	MINOR NORTH-SOUTH FOLDS (upright chevron style) axial-plane crenulation cleavage; crosscutting anisome	MINOR NORTH-SOUTH FOLDS Sleep east-dipping fracture west side zone only	NORTH-SOUTH FOLDS upright gentle variable plunge; steep cleavage
Devonian 410 Ma				MAIN ANTICLINE Develops over long period		
				Quartz veins minor		Unconformity
				Crystallization of micas and quartz; albite and garnet * (448-411 Ma)		
Silurian 438 Ma	Hinsburg Thrust Quartz veins in and across foliation	[F ₁] [S ₁]	Recrystallization of micas associated with S ₁ crenulation cleavage	[M ₁]	Weak recrystallization associated with cleavage	[M ₁]
Late Ordovician	Recrystallization micas, associated cleavage formation * (427-411 Ma)	[M ₁]	STEEP NORTH-SOUTH FOLDS (recined "Z" folds of bedding and quartz veins; crenulation cleavage (S ₁))	EAST-WEST CROSS FOLDS (recined, variable vergence "conjugate" folds, axial-plane crenulation cleavage)	STEEP NORTH-SOUTH FOLDS recined asymmetrical folds, axial-plane cleavage, shearing	[F ₁] [S ₁]
468 Ma	NORTH-SOUTH FOLDS Upright to inclined in west; variable plunge, axial-plane cleavage	[F ₁] [S ₁]	NORTH-SOUTH FOLDS Tight upright folds, S ₁ mostly crenulation or parallel to bedding; quartz veins mostly parallel to bedding, schistosity	Quartz segregation into S ₂ recrystallization of micas	NORTH-SOUTH FOLDS (?) Developed in Vermont (Cady et al., 1983)	[F ₁]
Middle Ordovician				BEDDING SCHISTOSITY [S ₀]		

prehnite-pumpellyite assemblages in the western Appalachian allochthons (Zen, 1974), and of crossitic amphibole in the Tibbit Hill schist of the Richmond area (Trzcinski, 1976) and glaucophane in mafic schist in north-central Vermont (Laird and Albee, 1981a), suggest the possibility that at least part of the Western Zone may have undergone high-pressure metamorphism (Sutter et al., 1985). Moreover, the common assemblage of chlorite, muscovite, albite, quartz, graphite and calcite in the Sutton schists, plus the association of greenstones (ophiolites?) and lack of granite intrusions, are all typical of high-pressure-series metamorphic zones (Miyashiro, 1973). Careful search, however, has failed to reveal any of the diagnostic high-pressure minerals in the report area.

Metamorphic mineral assemblages

The metamorphic assemblages in the Sutton area are of lower to upper greenschist facies, and they are dominated by a wide zone of albite schist coincident with the Sutton Anticline. The distribution of minerals is indicated in Figure 23. The Sutton schists contain variable proportions of quartz, muscovite, chlorite, albite, graphite, calcite, biotite, opaque minerals, tourmaline and zircon. Albite, tourmaline and opaques (titanomagnetite?) are

markedly porphyroblastic. The plagioclase is generally untwinned or broadly twinned albite; lamellar twinning is uncommon. Inclusions of groundmass in porphyroblasts, especially quartz, muscovite or graphite show important textural relationships. Biotite is only rarely developed, as wisps or patchy aggregates, probably because of the abundant graphite. No paragonite was detected; three samples of Sutton schist tested by X-ray diffraction indicate that the mica is muscovite with basal spacing from 9.962-9.965 (Å). The metabasites are laminated with albite- and chlorite-rich segregations in the schistosity. The albite is porphyroblastic, and the chlorite generally has an anomalous grey or khaki birefringence. Minor blue-green amphibole and green-brown biotite occur in metabasites as far west as the western edge of the albite porphyroblast zone as does the rare biotite in the pelites (Fig. 23). The high content of water and the $\text{Fe}_2\text{O}_3\text{-FeO}$ ratios for all of the metabasites (Table 2) compared to basalt is probably the result of metamorphism.

The highest grade zone in the core of the anticline is marked by the development of rare garnet and blue-green chloritoid in the semipelites, and by needles of colourless amphibole (tremolite?) along with chlorite, albite to andesine, epidote, calcite and green-brown biotite in metabasites. Similar zones of higher greenschist facies

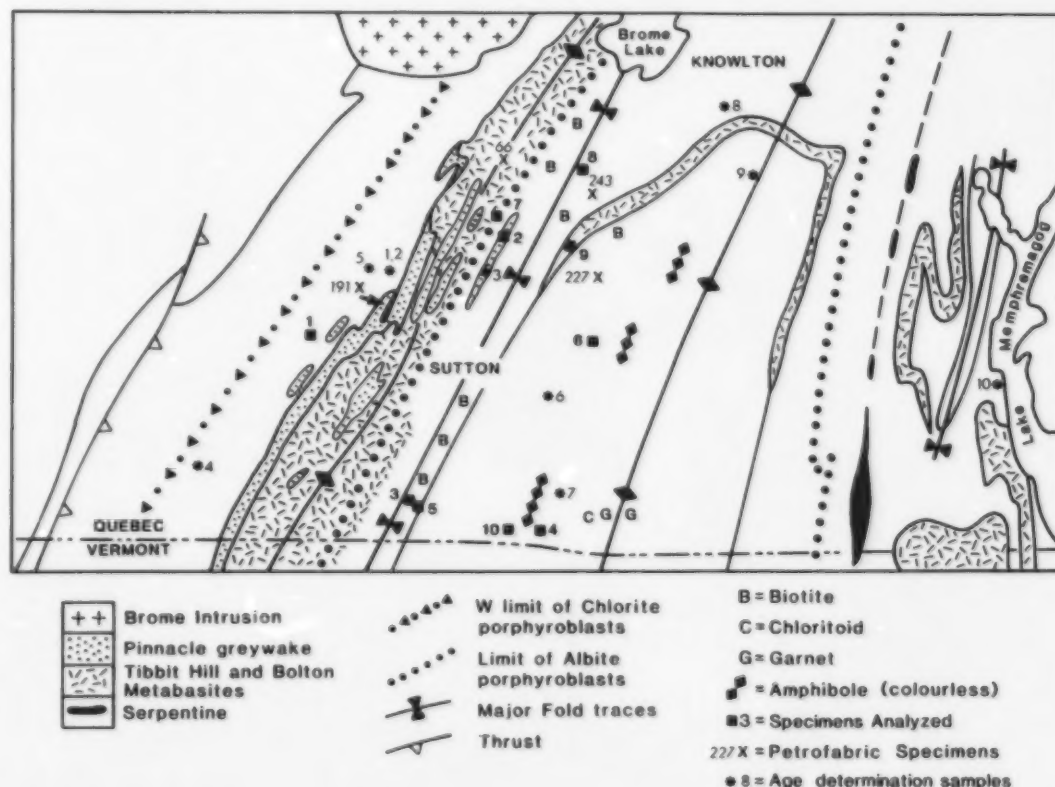


Figure 23. Map of the metamorphic zones with sample localities for petrofabric study, chemical analyses and age determinations. Additional information about samples for which age determinations were made is included in the Appendix.

occur along the axis of the Green Mountain Anticlinorium to the south, and the mineral assemblages there have been described in detail by Christman (1959), Albee (1968) and Laird and Albee (1981a,b). There is evidence for two phases of deformation and metamorphism, and discontinuously zoned garnet and amphibole porphyroblasts are reported (Albee, 1968; Laird and Albee, 1981b). In the Sutton area of this report, recrystallization during and after the late folding event appears to have been almost complete and, although there are abundant examples of relict microstructures, no zoned porphyroblasts or relict early minerals were found.

To the west, in the Enosburg Falls Anticlinorium, mineralogical changes are mild, even in pelitic rocks, and are associated with the development of a slaty-type cleavage (Fig. 24). Chlorite and sericite are developed throughout, but a wide zone in which sparse small (less than 1.5 mm) chlorite porphyroblasts (see Fig. 39), minute (0.2 mm) albite porphyroblasts and rare magnetite octahedra are developed, extends westward well into the lower Gilman phyllites (Fig. 23). Metamorphic tourmaline overgrowths on detrital grains occur as far west as the western edge of the report area (Rickard, 1964). The

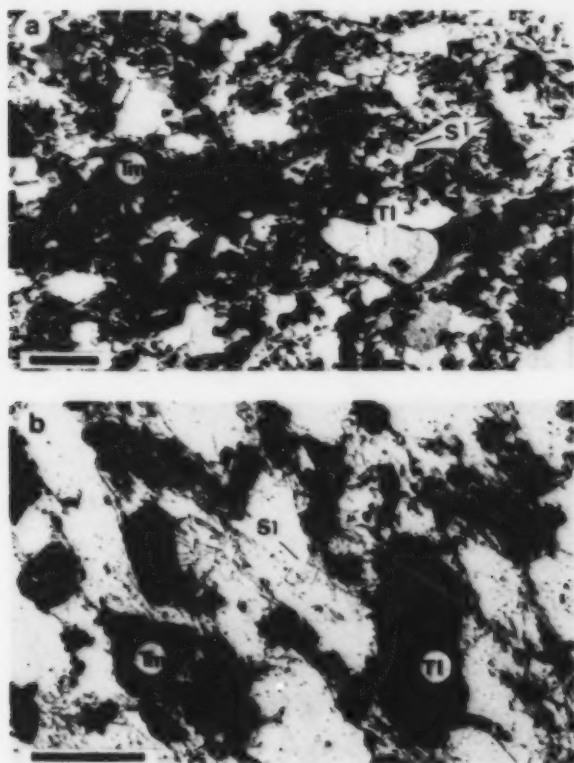


Figure 24. Incipient slaty cleavage in Pinnacle greywacke. a) A roughly aligned (S_1) muscovite and chlorite meshwork forms a matrix to detrital titanomagnetite (Tm), quartz and tourmaline (Ti). Scale 0.5 mm. Location—West Sutton (838982). b) Detail of mica alignment in S_1 , and metamorphic overgrowth of tourmaline (Ti). Scale 0.5 mm.

metabasites of the Tibbit Hill volcanics are strongly schistose and consist of a typical lower greenschist assemblage of chlorite, albite, epidote ($\pm$ zoisite or clinozoisite), calcite, sphene and opaques. Early formed twinned albite is abundant but long bladed crystals especially in flattened amygdules are in the process of breaking down to mosaics of untwinned crystals. Blue-green amphibole and green biotite were found at a few localities along the eastern edge of the outcrop belt. Metamorphic redistribution has resulted in abundant segregations and nodules of epidote (Pintson et al., 1985).

The eastern boundary of the albite zone coincides with the upper part of the Bonsecours schist. The eastern formations in the Mansonville area are of very low grade, although the sedimentary fabric is modified by the development of slaty cleavage (Fig. 25). The fine grained pelitic rocks contain abundant sericite and less common chlorite in clusters of minute flakes. The Bolton volcanics retain their original textures (Clark and Fairbairn, 1936), except at their margins where the pyroxenes have been altered to amphibole and the feldspars saussuritized or converted to albite; the matrix is a roughly schistose aggregate of albite, chlorite, tremolite, clinozoisite and sphene.

The mineralogy of the ophiolite and serpentinite suites to the north has been described by Laurent (1975); he concluded that low pressure and moderate temperature conditions prevailed during obduction. Clague et al. (1981) have calculated that the amphibolite beneath the Thetford Mines ophiolite was deformed at a pressure of about 5×10^5 kPa (5 kbars), and a temperature of 850°C was generated by frictional sliding during obduction.

Development of foliations

Severe textural changes are associated with the development of axial-plane foliations, and various stages in the processes are illustrated in Figures 24 to 29. Early clastic textures are partly preserved in the lowest grade rocks. The best examples are found in the heavy-mineral layers of the Pinnacle greywacke, where titanomagnetite (?), zircon, sphene and tourmaline still have rounded outlines

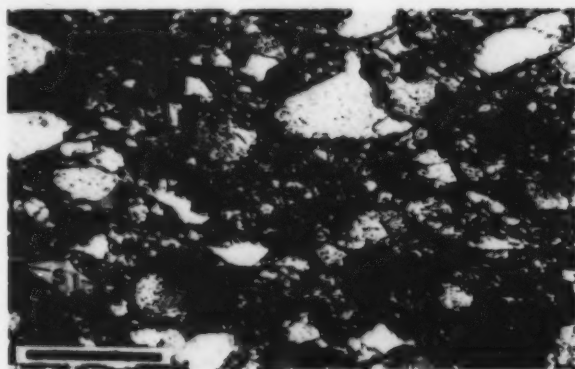


Figure 25. Slaty cleavage in Sweetsburg quartzite. Flattened detrital quartz grains in S_1 foliation. Scale 0.5 mm. Location—Highway 243, 2.5 km north of Mansonville (055941).

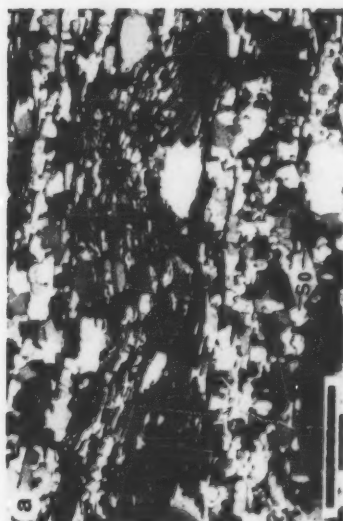


Figure 26. Photomicrographs illustrating quartz textures.

- a) Detrital quartz (Qz) grain reduced to a mosaic of strained subgrains in a pebbly layer of the Call Mill slate. Scale 0.5 mm. Location—1.5 km west of Mansville (863036).
- b) Mica "beards" (Mu) on quartz grain (Qz) in the Pinnacle greywacke. Scale 0.25 mm. Location—West Sutton (831991).

Figure 27. Photomicrographs illustrating textures of quartz segregations.

- a) Recrystallized polygonal quartz and mica-rich layers defining a bedding schistosity (S_0) in the Bonsecours schist. Scale 0.5 mm. Location—2.5 km southwest of Brome (897046).
- b) Polygonal quartz in folded veinlets parallel to bedding schistosity (S_0) in the Bonsecours schist. Scale 0.5 mm. Location—Townsville Road, 3.5 km east of Mansville (913035).

Figure 28. Photomicrographs illustrating early crenulation cleavage.

- a) S_{1x} in core of folded quartz veinlet in the Bonsecours schist. Scale 0.5 mm. Location—2.5 km southwest of Brome (899046).
- b) S_{1x} with differentiated muscovite-rich layers in Sallys Pond gneiss. Note that crenulations are defined by straight recrystallized mica flakes. Scale 0.5 mm. Location—Highway 105A, 1 km east of Glen Sutton (942906).

(Fig. 24). Quartz, too, preserves clastic outlines in the less deformed greywacke and some of the quartzite beds of the eastern outcrops of the Sweetsburg Formation (Fig. 25).

During the development of slaty cleavage, quartz and feldspar grains were flattened and strained, internal subgrain mosaics were developed and ultimately even at low metamorphic grade, irregular intergrowths and polygonal aggregates of smaller crystals resulted (Fig. 26 and 27). Most new growths of muscovite (sericite) and chlorite lie athwart the original bedding as an intergrain meshwork, and define the axial-plane foliation (Fig. 24). In places mica "beards" (Means, 1975) developed at the ends of quartz grains (Fig. 26b), and chlorite concentrates in small patchy porphyroblasts and individual flakes cut across the cleavage. Where bedding was originally thin laminae, as in the Sweetsburg and parts of the lower Gilman formations, the earliest cleavage developed as a close-spaced crenulation cleavage across the shaly foliation.

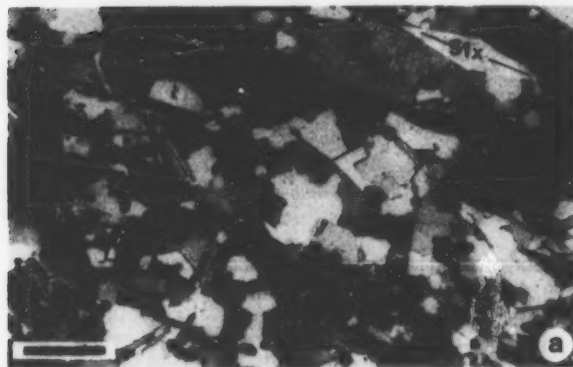
The lithological layering (bedding schistosity) in the Sutton schist is defined by a strong parallel orientation of muscovite and chlorite interlayered with quartz- and feldspar-rich layers (Fig. 27), so that all subsequent deformations produced a crenulation cleavage. Schists with the early (S_{1X}) crenulation cleavage are strongly segregated into micaceous (+ graphite) and quartzofeldspathic layers especially in the hinges of folds (Fig. 28). In the field these segregated laminae are similar to the bedding schistosity; they can be distinguished by the fact that they are not folded by F_{1X} folds, and under the microscope by the occurrence of obliquely-oriented micas, representing S_0 , across the quartzofeldspathic layers (Fig. 29) or by inclusions in albite crystals (see Fig. 35a).

These two schistosities (S_0 and S_{1X}) and the slaty cleavage (S_1) of the lower grade rocks were crenulated by the later F_2 folds, and locally a new axial-plane crenulation cleavage (S_2) developed (Fig. 30). Where growth of muscovite, chlorite, biotite or albite appears to be controlled by S_{1X} schistosity (Fig. 28 and 29) it is almost certainly mimetic, as relationships in nearby F_2 structures are identical (Fig. 31 and 32). New muscovite and albite grew across the earlier foliations in the axial-plane (S_2) direction (Fig. 30c, 31b and 32a). The micas defining the F_2 microfolds are usually straight strain-free flakes (Fig. 31a and 32b) indicating that recrystallization was generally syn- to post- F_2 deformation, a conclusion supported by the texture of porphyroblasts (see below). Mild effects of this late metamorphism extended as far west as Cowansville, as tourmaline overgrowths there include the S_2 foliation (Rickard, 1964). However, competent units such as the Pinnacle greywacke show only the S_1 foliation, and in a few places F_2 microfolds nearby are defined by bent micas. This and the evidence of folded and strained quartz veins indicate that locally, the greywackes may have been little influenced by recrystallization during the F_2 deformation.

Quartz veins

Several generations of quartz veins are developed. Many veins lie in the foliation planes and show complex relationships to folds (Fig. 33).

In the Sutton schists, quartz veinlets (a few millimetres to a few centimetres thick) segregated out parallel to the original bedding, probably aided by dewatering or devolatilization during the early development of the bedding schistosity. The extent of the quartz segregation is



a) Recrystallized muscovite defining F_{1X} crenulations and crenulation cleavage (S_{1X}) in Sallys Pond gneiss. Scale 0.5 mm. Location—Highway 105A, East Richford (902872).



b) Recrystallized muscovite mimetically controlled by F_{1X} crenulations in the Bonsecours schist. Scale 0.5 mm. Location—2 km northeast of Abercorn (858903).

Figure 29. Photomicrographs illustrating mineral textures in F_{1X} microfolds.

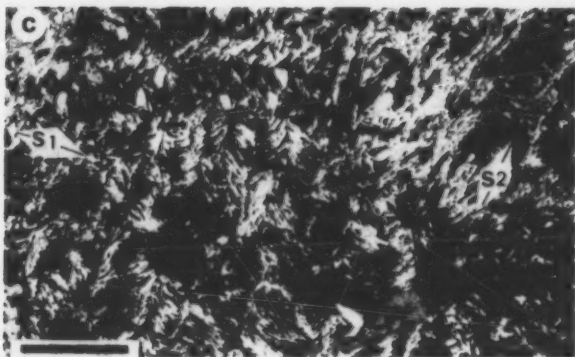
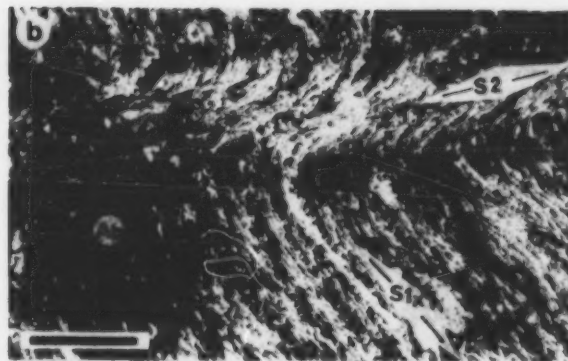
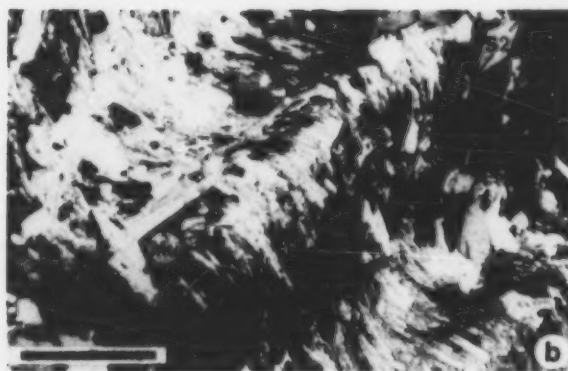
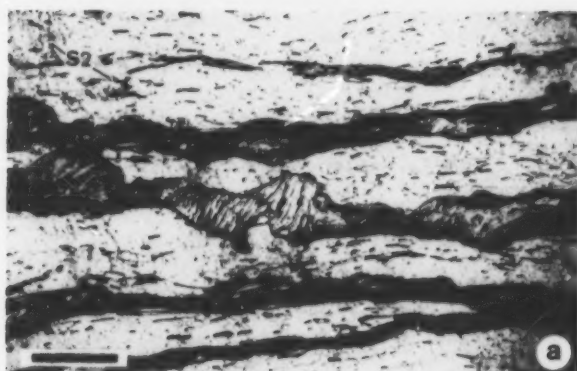


Figure 30. Photomicrographs illustrating superposed cleavages in the Bonsecours schists.

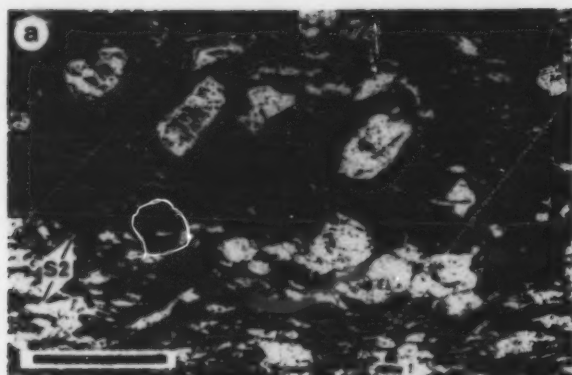
- a) Early crenulation cleavage (S_{1X}) crossed by late crenulation cleavage (S_2). Scale 0.5 mm. Location—0.5 km west of Brome (902070).
- b) F_2 crenulating S_{1X} crenulation cleavage. Note albite porphyroblasts growing across S_2 . Scale 0.5 mm. Location—1 km west of Sutton Junction (879021).
- c) Strong crenulation of S_{1X} crenulation cleavage. Note albite porphyroblasts aligned in the axial plane (S_2) of the late crenulations. Scale 0.5 mm. Location—2.5 km north of Sutton (873999).



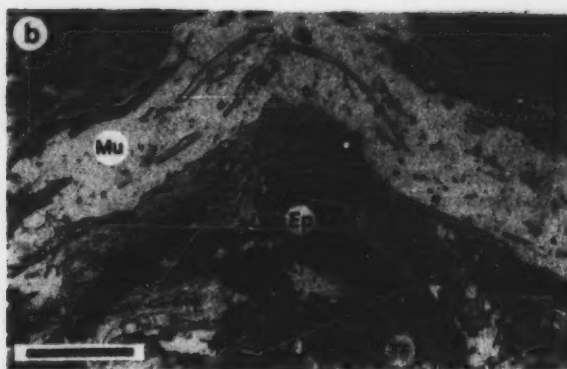
- a) Recrystallized muscovite in microfolds across late crenulation cleavage (S_2) in Bolton Glen schist. Scale 0.5 mm. Location—Townline Road, 3.9 km east of Mansville (916032).

- b) Recrystallized muscovite (Mu) in axial plane of late F_2 folds in Mohawk Lake schist. Scale 0.5 mm. Location—Morgans Road (885945).

Figure 31. Photomicrographs illustrating late crystallization of mica.



a) Albite porphyroblasts in axial planes of late F_2 folds, Bonsecours schist. Scale 0.5 mm. Location—1 km west of Sutton Junction (878021).



b) Recrystallized muscovite in straight strain-free flakes; epidote (Ep) and sphene (Ti) in late F_2 microfold, Bolton Glen schist. Scale 0.5 mm. Location—Townline Road, 4 km east of Mansville (916032).

Figure 32. Photomicrographs illustrating mineral growth in relation to late folds. Porphyroblastic growth is synchronous with or postdates late deformation.

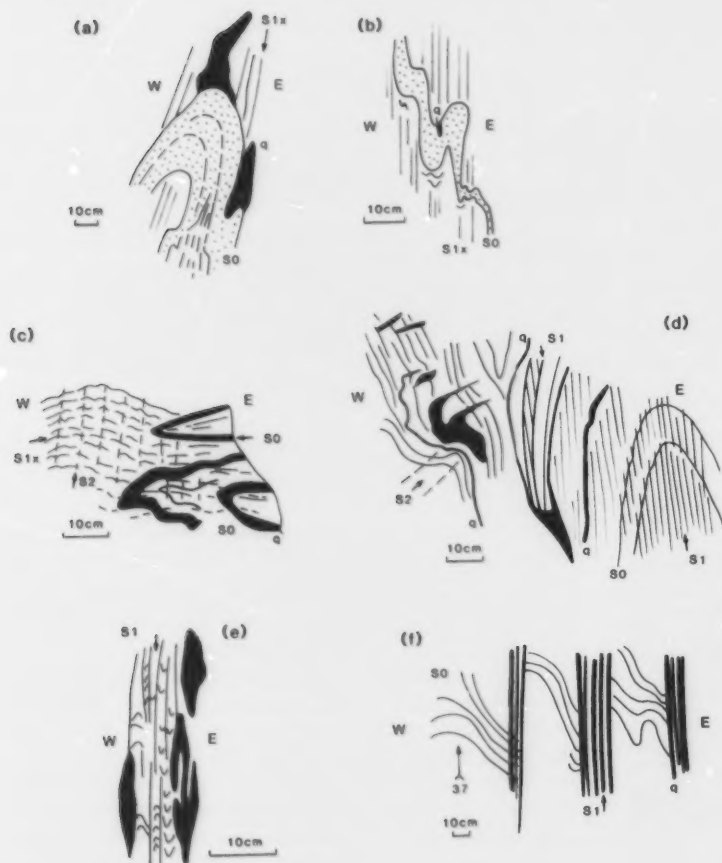


Figure 33. Quartz vein relationships to folds. (Quartz (q)) is black, quartzite dotted. a) Quartz gash veins in axial plane (S_{1x}) of F_{1X} fold. Quartzite layer in Mohawk Lake schist. Location—Muir Road, 7 km east of Mansville (949033). b) Quartz veinlet in axial plane (S_{1x}) of steeply-plunging fold in Sweetsburg quartzite. Location—1.5 km west of Peabody (049947). c) Folded (F_{1X}) quartz veinlets in bedding schistosity (S_0) in Bonsecours schist. Location—2 km southwest of Sutton (860963). d) Irregular quartz veinlets in bedding (S_0) and late cleavage (S_2) planes in lower Gilman phyllite. Location—2.5 km east-southeast of East Dunham (795967). e) Quartz veins in steep F_1 folds and axial plane (S_1) in Tibbit Hill schists. Location—Call Mill (871056). f) Quartz veinlets in shear zones parallel to S_1 axial plane in folded Pinnacle greywacke. Location—2.6 km west-northwest of Abercorn (819900).

indicated by two chemical analyses (Table 2, columns 5 and 6). Bonsecours phyllite (sample 5) from between quartz veinlets and Mohawk Lake graphitic albite schist (sample 6) have compositions unlike any sediment, being richer in aluminum and alkalis (cf. Pettijohn, 1975). The addition of about 40 percent silica to both samples converts the analyses to that of an average greywacke. These segregated veinlets are a prominent feature of the Sallys Pond, Mohawk Lake and Bonsecours schists where they are everywhere strongly folded by the early cross folds F_{1X} and shredded into quartz rods by the S_{1X} foliation (Fig. 15d,e and 33c). These rods mark the F_{1X} fold hinge direction. The folded quartz veins are generally unstrained, having been converted to polygonal mosaics (see Fig. 27b and 41a) by synchronous or later recrystallization.

Less commonly, coarser crystalline quartz lenses and veinlets occur in the axial plane (S_{1X}) of the folds in the Sutton schists (Fig. 33a,b). Both sets of quartz veins are folded and broken up in places by late (F_2) folds, and rare quartz veins lie parallel to their axial planes (S_2) (Fig. 33d). This may be younger quartz because thicker veins also cut obliquely across the S_2 direction and occupy joints.

In the Oak Hill rocks, quartz veinlets also parallel the bedding, especially in the Call Mill slate and Gilman phyllite, but more commonly they lie in the axial-plane (S_1) foliation (Fig. 33d,e,f). Extension gashes and irregular veins occur in the Pinnacle greywacke and both dolomite formations although, even there, most quartz lies in regular thin veinlets in the S_1 foliation (Fig. 8f and 33f). In the Tibbit Hill volcanics, quartz forms irregular veins and knots and also contains albite, epidote and chlorite segregations. These veins lie mostly in the S_1 or S_{1X} foliation (Fig. 33e). Early quartz veinlets are folded and cut by the S_1 slaty cleavage in the Sweetsburg and Stanbridge slates (Fig. 34); other veinlets lie in or slightly oblique to this foliation (S_1). Two generations of quartz are indicated, one before and one after the development of the slaty cleavage. Both sets are folded by the late crenulation cleavage (S_2) (Fig. 8e and 34).



Figure 34. Photomicrograph illustrating folded quartz veinlet in Vail slate. Bedding (S_0) and quartz veinlets are cut by slaty cleavage (S_1) and crenulation cleavage (S_2). Scale 1 mm. Location—Dunham (727998).

Porphyroblasts

The evidence is clear that porphyroblastic mineral growth occurred during and after the late F_2 deformation. New muscovite flakes grew across the S_{1X} foliation in the S_2 direction (Fig. 31b and 38) and a few schist layers are spangled with small muscovite porphyroblasts. Tourmaline prisms grew in irregular radiating aggregates but in a few places they are aligned in F_{1X} or F_2 crenulations. The small chlorite porphyroblasts that occur in the lower Gilman phyllite are commonly located in S_2 kink bands (see Fig. 39).

Only the albite porphyroblasts show useful inclusion textures as described by Vernon (1977). Albite occurs in stubby irregular to rounded subhedral prisms as individuals or in clusters; some have ragged terminations interfingering with micas (Fig. 35), others are curved following microfolds and in more quartzitic rocks large patches in the matrix extinguish together forming patchwork porphyroblasts (Fig. 36b). The albites are commonly poikiloblastic and include graphite, quartz, iron oxide, muscovite and more rarely epidote, apatite and biotite. The inclusions are smaller than the external matrix but commonly continue the matrix trends and preserve microfold textures (Fig. 35); these inclusion trails pass across twin-plane boundaries (Fig. 35c). The inclusion textures show that albite growth was post tectonic to both early (F_{1X}) and late (F_2) microfolds (Fig. 35b). Even where F_2 is weak the albites tend to have aligned prism faces or form aligned trains or lenticular veinlets in the local S_2 direction (Fig. 32a, 36a and 37). Rarely F_{1X} and F_2 microfolds are included in adjacent albite crystals or as refolded folds in the one crystal (Fig. 35b). It is not clear whether, or to what extent, albite crystallized during F_{1X} deformation, although a few of the included F_{1X} microfolds are more open than those in the adjacent matrix (Fig. 35a), suggesting that albite grew early and protected the folds from flattening. Graphite inclusions are in places restricted to core positions in albite but no clear examples of zoned crystals were found that might indicate two stages of growth.

The lack of early albite growth might be taken as evidence for a late soda metasomatism (Vernon, 1977). This is not supported by chemistry, however; samples of four different rock types from the porphyroblast zone (Table 2), albite gneiss (sample 4), graphitic albite schist (sample 6) and albitic greenschist (samples 9 and 10), contain less Na_2O than nonporphyroblastic equivalent rocks, namely greywacke (sample 2), graphitic phyllite (sample 3) and greenschist (sample 7).

Thus, although metamorphism accompanied each deformation stage, there is remarkably little evidence to indicate what the metamorphic state in the Sutton schists was during early deformations. It is probable that original sedimentary feldspar (albite, or possibly a more calcic variety) recrystallized and deformed several times to culminate in totally recrystallized albite porphyroblasts during a late tectonic to post tectonic thermal peak.

Similar albite porphyroblasts in the Dalradian schists of Scotland have been attributed to a hydrothermal retrogression of Barrovian biotite-garnet zone schists from upper to lower greenschist facies (Watkins, 1983).

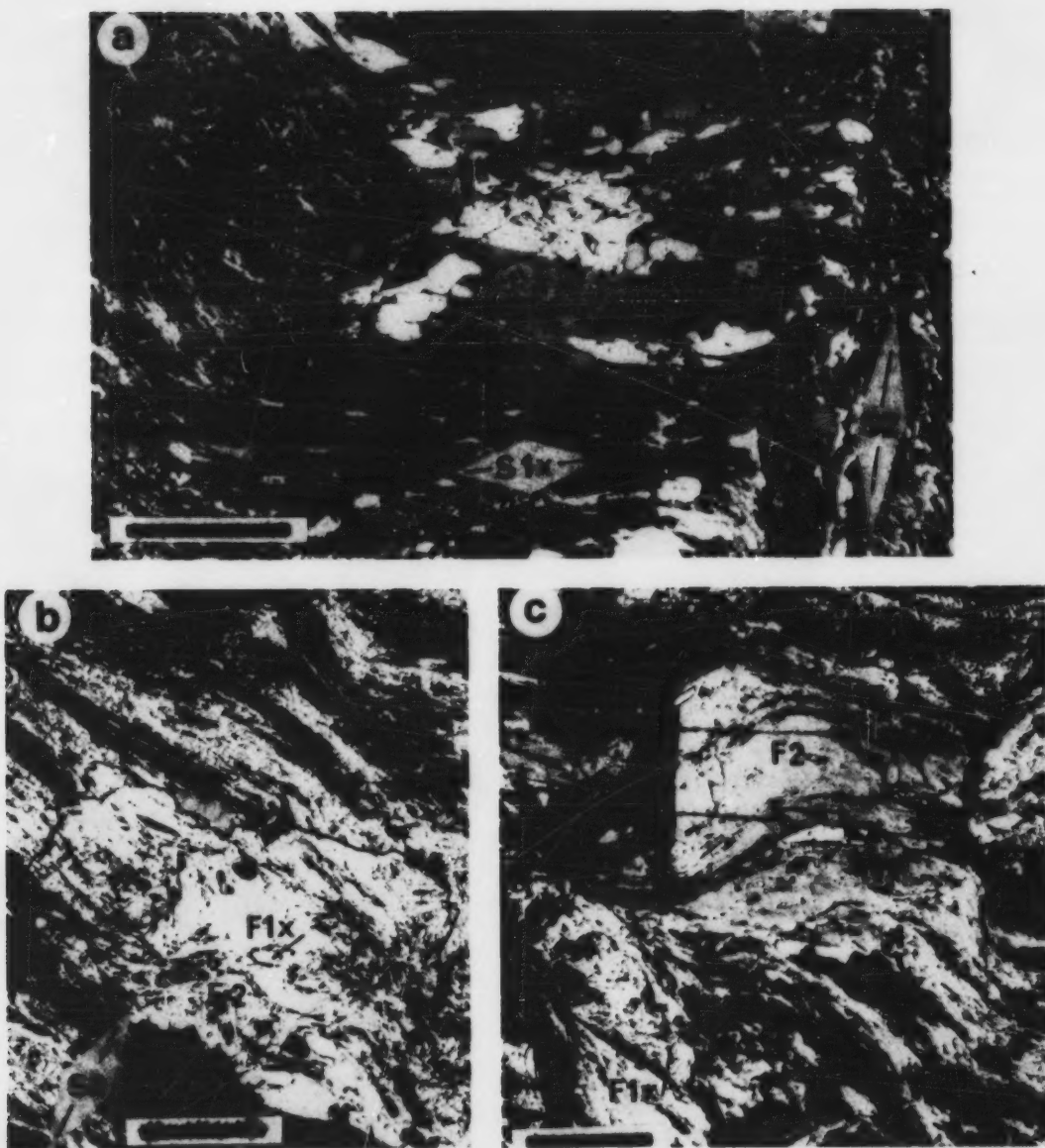


Figure 35. Photomicrographs illustrating the inclusion textures of albite porphyroblasts. Scale 0.5 mm.

- a)** Included relict open F_{1x} microfolds with axial-plane (S_{1x}) schistosity in the matrix, Bonsecours schist. Location—1 km west of Sutton Junction (878021).
- b)** Included F_{1x} isocline and F_2 microfold, illustrating refolding about S_2 axial plane, Mohawk Lake schist. Location—5 km east of Sutton (925978).
- c)** Twinned albite with included F_2 microfolds and F_{1x} microfolds in matrix. Location—5 km east of Sutton (925978).

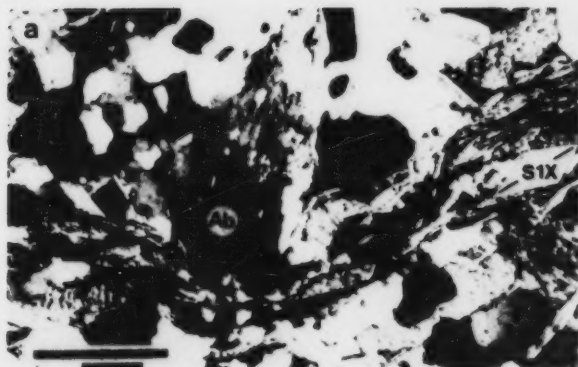


Figure 36. Photomicrographs illustrating albite porphyroblast textures in Sallys Pond gneiss. Scale 0.5 mm.

- a) Albite porphyroblast (Ab) distorting S_{1x} schistosity. Location—Highway 105A, 0.5 km north of East Richford (903874).
- b) Large patchwork porphyroblast of twinned albite with common extinction position. Location—Highway 105A, 0.5 km north of East Richford (903874).



Figure 37. Albite growth in late axial-plane direction in Sallys Pond gneiss.

- a) Trains of porphyroblasts aligned with the late crenulation cleavage (S_2) crossing the S_{1x} schistosity. Scale 1 mm. Location—Highway 243 near Sallys Pond (008061).
- b) Detail of a). Scale 0.6 mm.

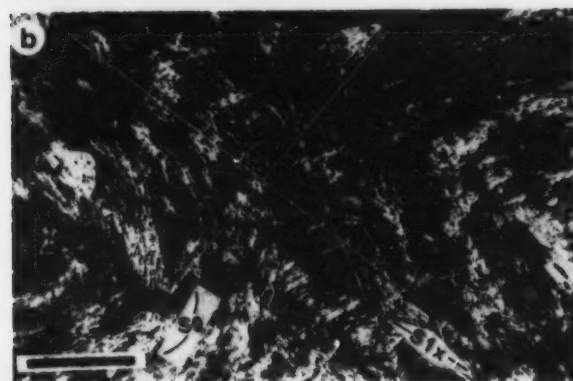
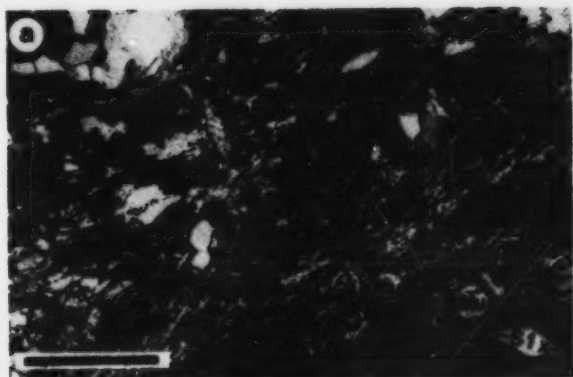


Figure 38. Mica growth in late axial-plane direction in Sallys Pond gneiss. Scale 0.5 mm.

- a) Muscovite (Mu) porphyroblasts oriented across early schistosity (S_{1x}).
b) Muscovite (Mu) porphyroblasts in S_2 direction lying across S_{1x} direction in F_{1x} microfolds. Location a and b—Highway 243 near Sallys Pond (008061).



Figure 39. Chlorite (Ch) and small pinhead albite porphyroblasts in S_2 kink bands in West Sutton slate. Scale 0.5 mm. Location—Nevin Road, 1 km northwest of West Sutton (828990).

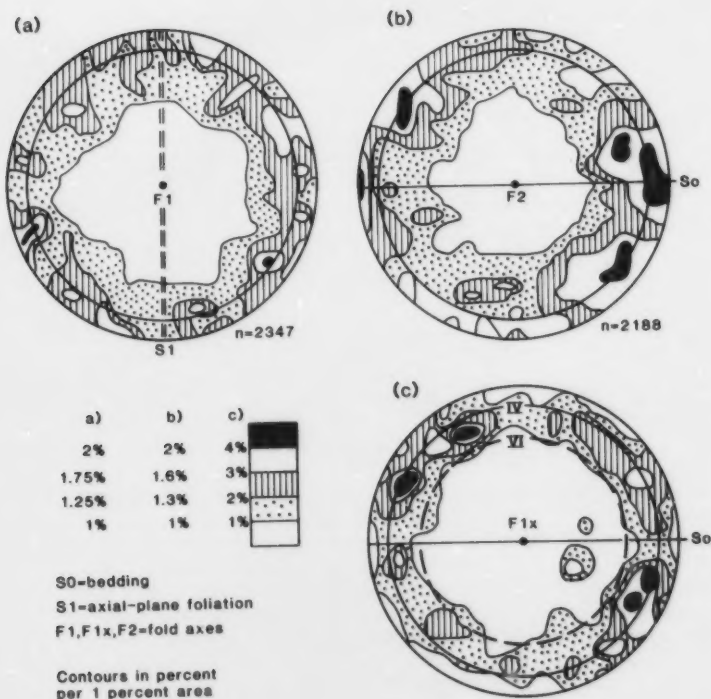


Figure 40. Summary petrofabric diagrams. Quartz c-axes are plotted on the lower hemisphere of equal-area Schmidt nets. a) Eight quartzites in the Oak Hill Group, possibly from the Pinnacle greywacke or upper Gilman quartzite. b) Seven specimens of Sutton schists. c) One representative sample of Sweetsburg quartzite, at Mansonville (n not indicated). Type IV and VI small-circle girdles are indicated. a) and b) are from Fairbairn (1935, Fig. 1) with permission of the University of Chicago Press. c) is from Fairbairn (1939, Fig. 9b) with permission of the Geological Society of America.

The process causes segregation of quartz and depletion of graphite at a grade insufficient to stabilize biotite. In contrast, in the Sutton schists quartz segregates early (pre- F_1), albite porphyroblasts grow late (post- F_2) and graphite is still abundant. Bell and Brothers (1985) described similar albite porphyroblast textures in the blue schists of New Caledonia and attributed their growth to replacement of omphacite during depressurization and carbonization associated with late-stage compressive deformation and thermal metamorphism. From the evidence in Vermont (Laird and Albee, 1981b; Laird et al., 1984) it is possible that there was an earlier higher grade metamorphic assemblage in the Sutton Anticline, but the only remaining evidence for this is the local isolated layers of garnet-chloritoid schist. The few garnet porphyroblasts are rimmed by chlorite which may indicate that they were formed early and later partially retrogressed. Albee (1968) was able to support this conclusion with the evidence of smaller unaltered early garnet crystals included in albite porphyroblasts, but no such garnets were found in this study.

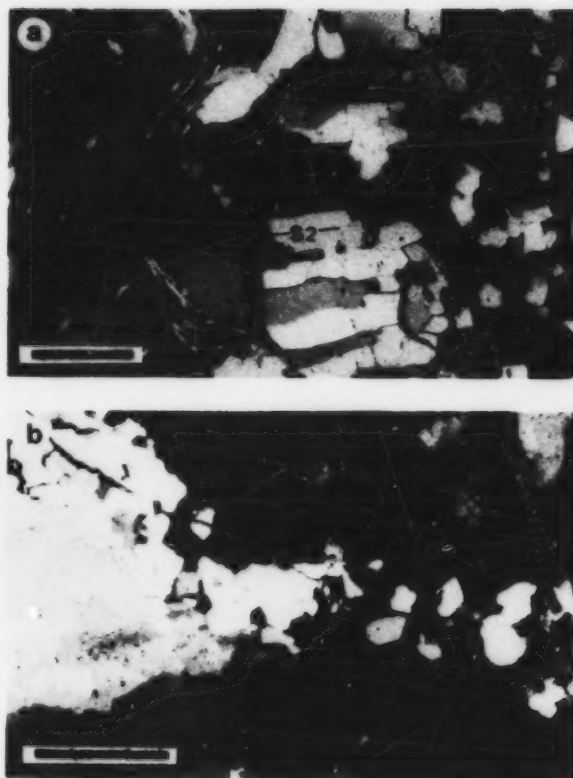


Figure 41. Photomicrographs illustrating texture of quartz.

- a) Deformation bands in quartz at hinge of F_2 microfold, Bonsecours schist. Scale 0.25 mm. Location—1.5 km north of Sutton (874988).
- b) New polygonal grains along junction of old deformed crystals in folded quartz vein in Pinnacle greywacke. Scale 0.5 mm. Location—2.1 km southwest of Brome (890059).

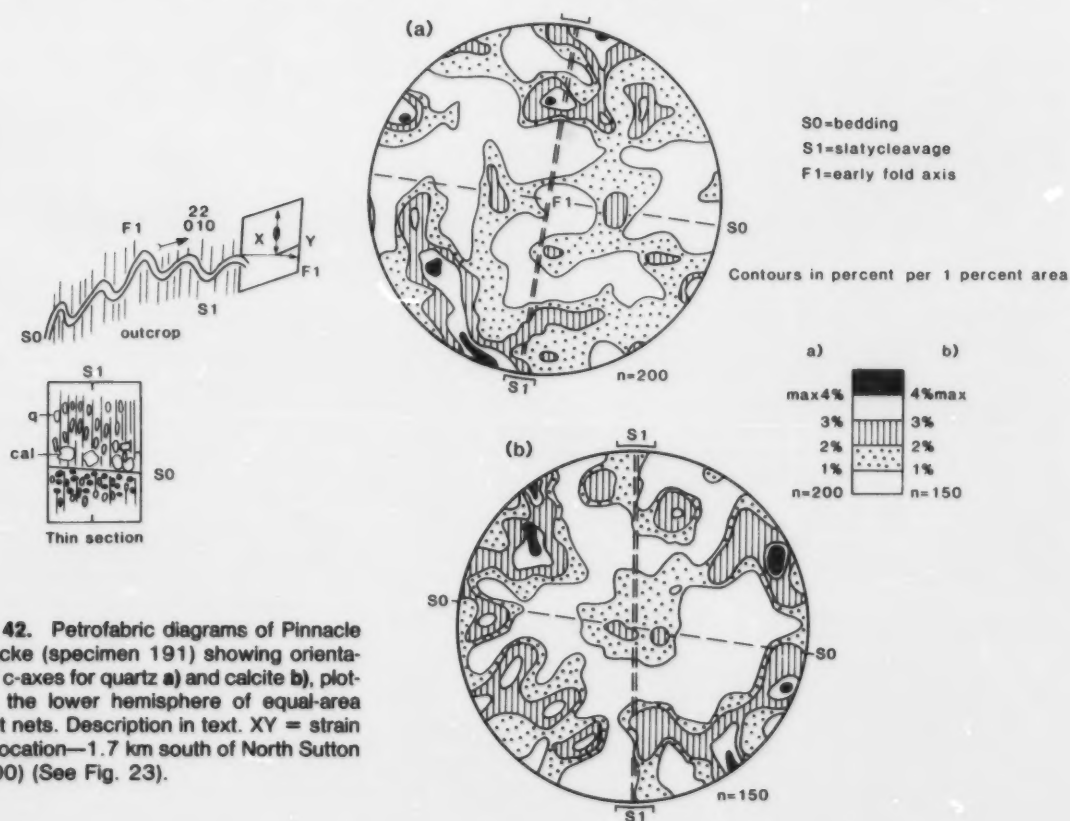


Figure 42. Petrofabric diagrams of Pinnacle greywacke (specimen 191) showing orientations of c-axes for quartz a) and calcite b), plotted on the lower hemisphere of equal-area Schmidt nets. Description in text. XY = strain axes. Location—1.7 km south of North Sutton (839000) (See Fig. 23).

Petrofabrics

In view of the important role of quartz segregation and the evidence for late recrystallization of feldspar and mica it is appropriate to examine the relationship between the growth of quartz and the microstructures. Fairbairn (1935, 1939), in a now classic petrofabric study, examined quartzites from the Oak Hill Group (possibly Pinnacle greywacke or upper Gilman quartzite), Sutton schists and the Mansonville Group (Sweetsburg quartzite). The c-axis patterns he obtained are characterized by the development of 50-70 degree small-circle distributions about the fold axis, and pseudo-orthorhombic symmetry (Fig. 40). In the present study the quartz fabrics of three specimens were determined specifically to evaluate the influences of folding, especially cross folding (F_{1X}), and post tectonic mineral growth associated with F_2 folding.

The general shape of quartz crystals in these rocks suggests that they have been extensively recrystallized since most have straight crystal faces and polygonal shapes (see Fig. 24 to 29). Where micas are abundant the quartz forms elongated "bricks" bounded by mica (Fig. 27a). Nevertheless, the effect of plastic strain is indicated by irregular extinction, subgrain development and deformation bands; in places, the bands develop into subgrains elongated parallel to the late (S_2) axial plane (Fig. 41a). Larger crystals are highly strained (Fig. 26a and 41b). The early quartz-vein fabric has been strongly deformed by late (F_2) folding, but the deformational features have not been entirely relieved by recrystallization.

In a folded and cleaved (S_1) calcareous greywacke that has elongated slightly strained quartz crystals set in a meshwork of muscovite and chlorite, there is a moderate development of new polygonal quartz crystals, mostly

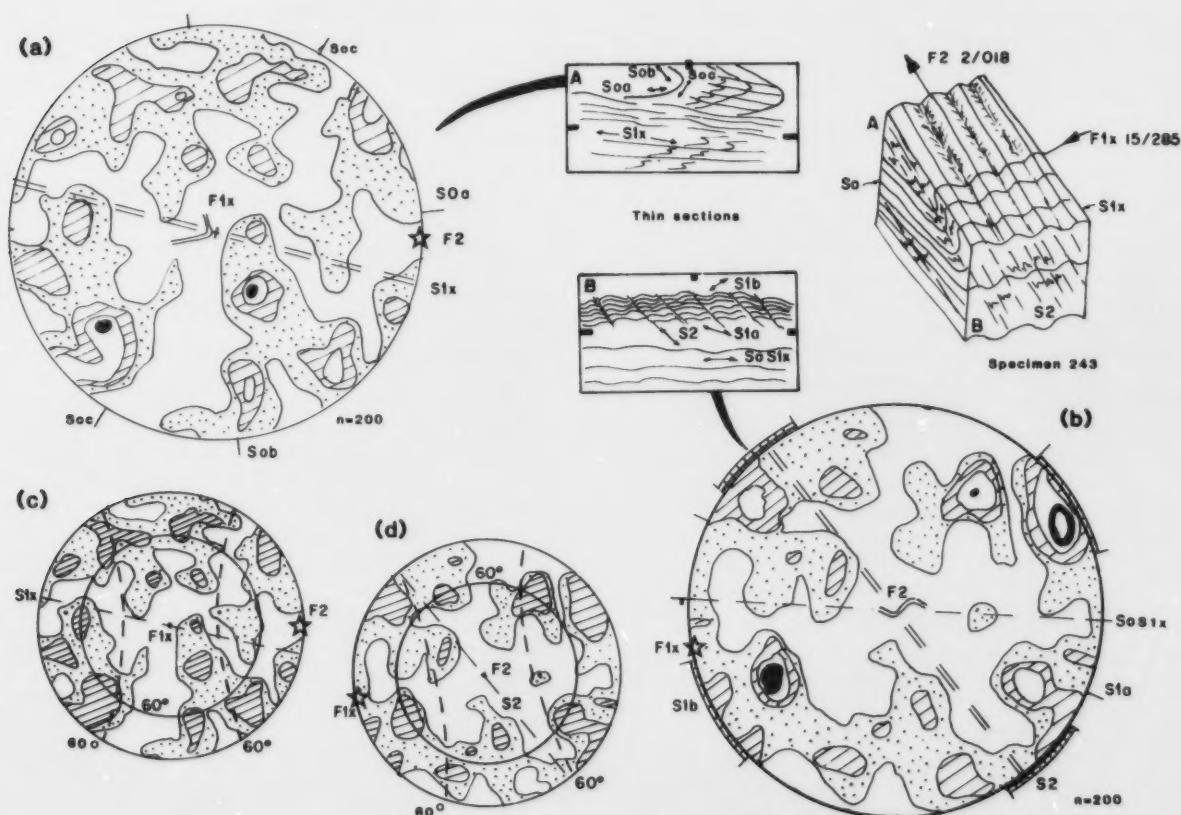


Figure 43. Petrofabric diagram for Bonsecours schist (Specimen 243) showing orientation of c-axes for quartz plotted on the lower hemisphere of equal-area Schmidt nets. a) Thin section cut perpendicular to F_{1X} fold. b) Thin section cut perpendicular to F_2 fold. Insets show structure of specimens and thin sections. c) and d) Interpretations of a) and b) showing type IV 60° small-circle girdles about F_{1X} and F_2 . Description in text. Contours are in percentage per 1 percent area (1%, 2%, 3%, 4%, 5%). S_0 = bedding; S_{0abc} = folded portions of bedding; S_{1X} = early crenulation cleavage; S_{1ab} = folded portions of S_{1X} ; S_2 = late crenulation cleavage; F_{1X} and F_2 corresponding fold axes. Location—3.5 km southeast of Brome (925041) (see Fig. 23).

too small to measure. Some larger calcite plates occur, especially near the detrital layers, and a few are twinned. The quartz and calcite show similar small-circle fabric patterns about the fold axes with incomplete cross girdles (Fig. 42a,b).

The specimens of Bonsecours and Bolton Glen schists have equally developed early cross folds (F_{1X}) and late folds (F_2) even on a microscopic scale; sections cut perpendicular to each fold axis were studied (Fig. 43 and 44). In the Bonsecours schist, mica flakes are bent and there are only a few small albite porphyroblasts. Quartz occurs as intergrowths of larger strained crystals and smaller new ones. The Bolton Glen schist is more recrystallized, micas are straight and most of the quartz is in smaller brick-like crystals with only a few larger strained remnant crystals. The sections perpendicular to F_{1X} (Fig. 43a and 44a) show diffuse distributions of quartz c-axes weakly symmetrical about the S_{1X} foliation. The sections cut perpendicular to F_2 (Fig. 43b and 44b) are somewhat different from each other. The distribution of quartz c-axes for the Bonsecours schist (Fig. 43b) closely resembles Fairbairn's (1935, 1939) small-circle girdle pattern although it lacks the concentration in the bedding plane

(S_0) (cf. Fig. 40b and 43b). The Bolton Glen schist yields a diffuse cross-girdle pattern with a strong concentration in the S_2 foliation (Fig. 44b).

The interpretation of these fabrics is difficult; clearly the inhomogeneities of rock structure are reflected in the mineral fabrics. The tendency for maxima to lie inside the perimeter of the nets in all the plots (Fig. 42, 43 and 44) yields small-circle girdles. An "orthorhombic" distribution of maxima in these 60-70 degree small-circle girdles has been commonly reported in other studies (Hara, 1971; Hobbs et al., 1976; Majoribanks, 1976). The zero to low concentrations of c-axes normal to the axial foliations and parallel to the fold axes suggest by analogy with computer simulations (Lister and Hobbs, 1980) and examples where strain has been measured (Majoribanks, 1976), that these rocks were deformed by flattening. Where strain axes can be deduced, in the Oak Hill rocks, the extension directions (X) lie in the axial foliation sub-perpendicular to the fold axes (Y) (Fig. 42). There is, however, little indication of the expected 30 degree small circles (Hara, 1971) about the normal to the foliation in any of the Sutton diagrams.

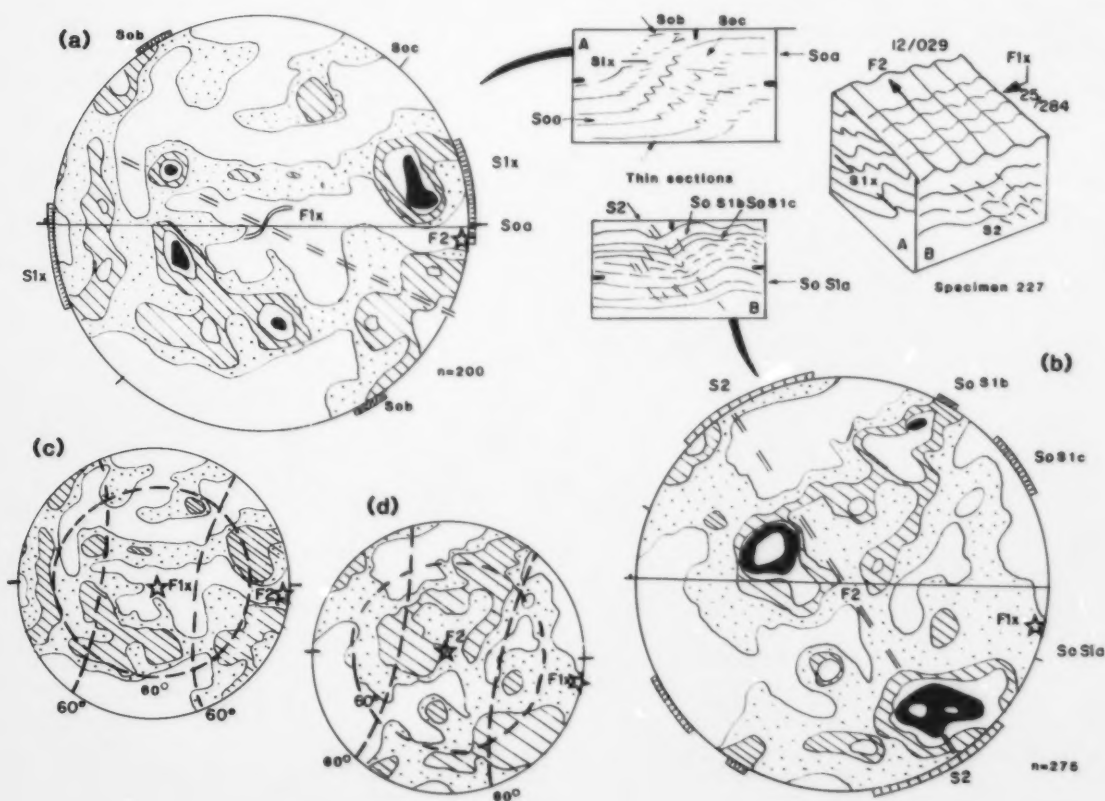


Figure 44. Petrofabric diagram for Bolton Glen Schist (specimen 227) showing orientation of c-axes for quartz. **a)** Thin section cut perpendicular to F_{1X} fold. **b)** Thin section cut perpendicular to F_2 fold. Insets show structure of specimen and thin sections. **c)** and **d)** Interpretations of **a)** and **b)**, description in text. Contours and symbols as for Figure 43. Location—Townline Road, 3.8 km east of Mansville (916032) (see Fig. 23).

The tendency to diffuse cross-girdle patterns in the sections of Bonsecours and Bolton Glen schists (Fig. 43 and 44) are perhaps best explained as partial girdles about both the F_{1X} and F_2 fold axes (Fig. 43c,d and 44c,d), thereby implying that early fabric elements are preserved. Lister (1977) commented that such "orthorhombic" girdles may result from late-stage coaxial plane-strain increments superimposed on earlier probably noncoaxial fabrics. In spite of clear evidence from metamorphic mineral textures that muscovite, albite and quartz recrystallized during and after late (F_2) folding, relict quartz fabrics related to the early F_{1X} axes appear to have been preserved. Either the quartz has not recrystallized completely or there is a strong mimetic control by elements of the early microfolds, or a host-grain control on orientation operated during the late recrystallization.

A problem with this interpretation is that the effects of F_2 deformation appear to be mild outside the Central Zone of Sutton schists, although the early and late deformations are approximately coaxial in the Oak Hill rocks. In the Pinnacle greywacke specimen (Fig. 42) there is no sign of late structures, yet the quartz fabric is similar to that of the Sutton schists (cf. Fig. 42, 43 and 44). Similarly the Oak Hill and Sweetsburg quartzites studied by Fairbairn (1935, 1939) are from areas in which F_2 effects are minimal and different single phases of folding are dominant—the Oak Hill quartzite has a gentle F_1 plunge and the Sweetsburg quartzite a steep F_{1X} plunge—yet the quartz fabrics are similar (Fig. 40). Thus it appears that similar "orthorhombic" small-circle girdles may have developed during all three stages of deformation.

AGE OF DEFORMATION

There has been much debate in the past over the age of the deformations affecting the Cambro-Ordovician rocks in the western part of the Appalachians. From stratigraphic evidence, the latest emplacement of the Taconic and related nappes is clearly post-Middle Ordovician (St. Julien and Hubert, 1975), but the folds and thrusts affecting them and the Green Mountain-Sutton Mountain Anticlinorium have been attributed to Taconic (Ordovician) or Acadian (Devonian) deformations by different workers (Rickard, 1965; Beland, 1967; Albee, 1968; Cady, 1969; Rodgers, 1971). There is now general acceptance that deformation and metamorphism of Taconic age occurred along the western margin of the Appalachians prior to the Acadian Orogeny, and that the Devonian movements had little effect to the west of the eastern flank of the Green Mountain-Sutton Mountain Anticlinorium (Zen, 1972; Carrara and Fyson, 1973; St. Julien and Hubert, 1975; Keppie et al., 1983; Stanley and Ratcliffe, 1985; Sutter et al., 1985). Briefly, the evidence is:

1. the unconformity between the steeply folded Cambro-Ordovician and the gently folded Silurian rocks at Sargent Bay (Rickard, 1965, present work; Lamothe, 1981);

2. retrogression along the western side of the boundary between eastern and western metamorphic zones in Vermont (Albee, 1968; Lanphere and Albee, 1974; Laird and Albee, 1981b; Karabinos, 1984; Laird et al., 1984) and a truncation of Taconic isotherms by the Acadian ones in Massachusetts (Sutter et al., 1985);
3. a western grouping of older (471–446 Ma) and eastern grouping of younger (375–335 Ma) radiometric dates* (mostly K-Ar or $^{40}\text{Ar}/^{39}\text{Ar}$), with a boundary along the eastern flank of the Green Mountain Anticlinorium (Harper, 1968; Lanphere and Albee, 1974; Laird et al., 1984; Sutter et al., 1985).

In the Sutton area, an attempt to date the two major fold episodes determined there by selecting samples on the basis of the S_1 and S_2 foliation textures (Lowdon et al., 1963; Rickard, 1965) yielded dates of 427 Ma for F_1 in the Pinnacle greywacke and 448 Ma for F_2 in the Sutton schist. It was concluded that both sets of folds were formed in the Taconic Orogeny (Rickard, 1965). Knight (1976) extended this work and obtained dates of 421 Ma for Sutton schists and 411 to 382 Ma for rocks of the Oak Hill Group. All dates in the report area are listed in the Appendix. The Pinnacle greywacke was resampled in 1976 and yielded dates of 427 Ma and 390 Ma. From new localities east of Knowlton, Bolton Glen schist, with well-developed F_{1X} and S_{1X} structures, yielded dates of 367 (K-Ar) and 411 (Rb-Sr) Ma, and Sallys Pond schist with well-developed F_2 and S_2 structures were dated at 380 Ma (Wanless et al., 1979). The K-Ar dates for the Sutton schists are essentially the same, and they confirm that rocks with early structures do not necessarily preserve early mineral growths. Nevertheless, it is clear that dates significantly older than Middle Devonian (Acadian) are preserved in the Sutton schists and Oak Hill Group, and that there is a spread of dates up to the Acadian values (375–335 Ma) as in areas to the north and south. Other older dates have been determined from nearby localities: in northern Vermont, 446 Ma and 430 Ma from the Hazens Notch schist (Harper, 1968) and 439 Ma from amphiboles in the Tibbit Hill schist (Laird et al., 1984); and to the north of the report area, 449 Ma from the Caldwell quartzite and 421 Ma from schists of the Rosaire Group (St. Julien and Hubert, 1975).

Interpretation of these dates is still equivocal. The spread of dates could be due to slow cooling after the Taconic events (Sutter et al., 1985), to superposition of Acadian deformation and metamorphism (Dennis, 1968; Cady, 1969) or to static degassing during Acadian events to the east. The evidence of metamorphic fragments in Late Ordovician sediments (Beaulieu et al., 1980), the likelihood of updoming and nappe removal, the unconformity below the Siluro-Devonian and the lack of granites all support a rapid uplift and cooling for the Sutton Mountain axis. Nevertheless, a spread of ages from Middle Ordovician to Late Silurian is to be expected with multiple deformation and variable rates of uplift and cooling. The fact that rocks in the Oak Hill Group containing only the older S_1 foliation (Pinnacle greywacke)

* All dates have been corrected to the latest International Standards (Steiger and Jager, 1977)

give older dates (427 Ma and 411 Ma) than rocks generally affected by the late S_2 cleavage (West Sutton slate, 372 Ma, and lower Gilman phyllite, 388 Ma), and that some of the Sutton schists strongly affected by the F_2 structures give the same dates (380 Ma, 367 Ma), suggests that separate dynamothermal events have operated. Moreover, the St. Daniel slates at Vale Perkins, close to the Sargent Bay Syncline have been dated at 365 Ma (Harper, 1968) and a late cleavage zone cutting the Granby nappe has yielded dates of 359 to 383 Ma (St. Julien and Hubert, 1975). These younger dates support the interpretation that the late deformation (D_2) and the albite porphyroblast metamorphism may be superposed on Taconic structures during the Acadian (Table 3). St. Julien et al. (1983) also interpreted the late folds of the Notre Dame Anticlinorium to the northeast of this area as Acadian.

The preservation of older dates (448 Ma, 421 Ma, 411 Ma) and relict fabrics in the Sutton schists presumably indicate that the schists were not totally re-equilibrated during metamorphism in spite of textural evidence for post- F_2 recrystallization. Alternatively, the older dates may represent Taconic cooling ages after the late deformation and metamorphism, and the younger dates may result from passive Acadian degassing of variable intensity and westward extent.

These various possibilities cannot be satisfactorily resolved at present, and the precise timing of events listed in Table 3 is difficult. The following age limits are given by stratigraphic and structural relationships in the nearby regions: i) Early Ordovician dates for the ophiolite emplacement (Laurent, 1975; Clague et al., 1981); ii) early F_1 and late F_2 structures deform the early Middle Ordovician slates of the Stanbridge Formation at Cowansville; iii) the 427 to 411 Ma cooling dates for the S_1 cleavage in the Pinnacle greywacke; iv) what appear to be F_{1X} folds and S_{1X} cleavage in the St. Daniel slate are cut unconformably by the Late Silurian Glenbrooke Group; and v) the age of the Acadian deformation at Lake Memphremagog has been determined by Boucot (1968) on stratigraphic grounds to be Eifelian (about 385 Ma).

It is concluded that three deformational stages developed in the geoclinal welt as strain thresholds were reached, depending on local conditions of ductility. In an essentially continuous process of obduction-imposed collapse, local early nappe development was followed by F_1 and F_{1X} folding during the Late Ordovician to Middle Silurian. The age of the late folding (F_2) in the Sutton area depends entirely on the interpretation of radiometric dates. Late folding probably occurred during the Acadian Orogeny when the Taconic "basement" made adjustments in which a widespread crenulation cleavage developed and the Sutton Mountain-Green Mountain Anticlinorium underwent reheating.

HISTORY OF DEFORMATION — CONCLUSIONS

The important stages in the deformational history are illustrated in Figures 45 and 46. In these conclusions, the deformational events are related to the development of the orogen in southern Canada and northern Vermont in terms of the current plate tectonic model (Laurent, 1975; St. Julien and Hubert, 1975; Poole, 1976; Osberg, 1978; Schenk, 1978; Williams, 1979; 1984; St. Julien et al., 1983; Stanley and Ratcliffe, 1985).

From late Precambrian until the Early Ordovician the paleogeography was that of a miogeoclinal or continental-rise prism (Williams, 1984) as shown in Figures 4 and 45a. The early Paleozoic continental margin of the Iapetus (proto-Atlantic) Ocean (Wilson, 1966) has been detected in several places along the northern Appalachian chain (Dewey, 1969; Williams and Stevens, 1974; Kumarapeli, 1985). The continental edge is marked by a sharp gradient in the gravity anomalies (Rabinowitz and LaBrecque, 1977; Cook and Oliver, 1981), an elongated wedge of arkosic sediments associated with basic volcanic rocks (in this case the alkaline metabasalts of the Tibbit Hill) and the close association of a carbonate-bank scarp (Rodgers, 1968; Bird and Dewey, 1970). The continental edge probably originated as a rift fracture in older basement (Rankin, 1976; Kumarapeli, 1985), and Thomas (1977) has related the salients in the continent margin and the accumulation of thick clastic wedges to the transform offsets of the original fracture (Fig. 46a). Cambrian alkaline igneous activity at about 574 Ma on the Precambrian Shield northwest of the Appalachian Orogen was associated with the rift system (Doig, 1970), but the oceanic margin must be older as sediments of late Precambrian and Early Cambrian age accumulated there (Fig. 45a). A continent-edge upwarp probably developed at the fractured margin (Rabinowitz and LaBrecque, 1977) giving rise to a geanticlinal structure similar to the present Atlantic margin; this situation also occurs to the north (Williams and Stevens, 1974). By Late Cambrian time, fine muds in a broad sag basin had developed across the continental-rise prism and continental edge.

Major deformation commenced in the Early Ordovician when obduction of serpentinite occurred (Fig. 45b) and the Atlantic-type margin was converted into a fringing island-arc system by the development of a small ocean basin and volcanic arc in the Middle Ordovician (St. Julien and Hubert, 1975; Doolan et al., 1982; Bradley and Kusky, 1986). The necessary subduction zone accompanying this obduction has been variously interpreted to dip westward (Bird and Dewey, 1970; Poole, 1976; Osberg, 1978) or eastward (Williams and Stevens, 1974; Hatch, 1982; Bradley, 1983; Williams, 1984; Bradley and Kusky, 1986; Coish et al., 1986). An easterly dip is more likely to explain the obduction and lack of Ordovician igneous activity in the continental-rise prism (Williams and Stevens, 1974).

These initial movements were accompanied by a rise of temperature in the thick continental-rise prism, and some thrusting involving oceanic basement slices (now serpentinite lenses) probably occurred (Fig. 45b); quartz segregated into a schistosity developing parallel to bedding, possibly under the influence of thrust shearing (Mataur, 1975). As compression increased, some thrust telescoping may have occurred (Bird and Dewey, 1970), and Ordovician strata below the obduction zone probably piled up over the continental-rise prism (Stanley et al., 1984; Coish et al., 1986). This process of nappe loading could explain the isolated occurrences of high-pressure facies metamorphics in the schist belt (Zen, 1974; Trzcien-ski, 1976; Laird et al., 1984; Sutter et al., 1985). The dehydration of such an over-thickened pile undergoing metamorphism could generate sufficient fluid pressure to initiate nappe sliding (Ayrton, 1980). As nappes slid or were pushed westward to form the Stanbridge and Granby nappes (St. Julien and Hubert, 1975; St. Julien

et al., 1983) or Taconic nappes (Cady, 1968a; Zen, 1972; Stanley and Ratcliffe, 1985), the continental-rise prism began to dome upward as it was compressed against the edge of the continental crust (Fig. 45b). As in the southern Appalachians (Wehr and Glover, 1985), this hinge zone had a marked effect on the developing structures; the underlying early paleogeographic regimes, continental shelf, continental-rise prism and ocean floor control different structural zones (Fig. 4), and the boundaries between them are marked by significant strain discontinuities. Brown (1972) has reported a similar situation in southern New Brunswick where zones of different structural style were dependent upon stress concentrations at the shelf-slope margin.

Continuation of compression in the Middle to Late Ordovician caused the Oak Hill strata on the continental shelf to crumple and slide westward on the Hinesburg Thrust as a para-autochthonous sheet; a fanning system

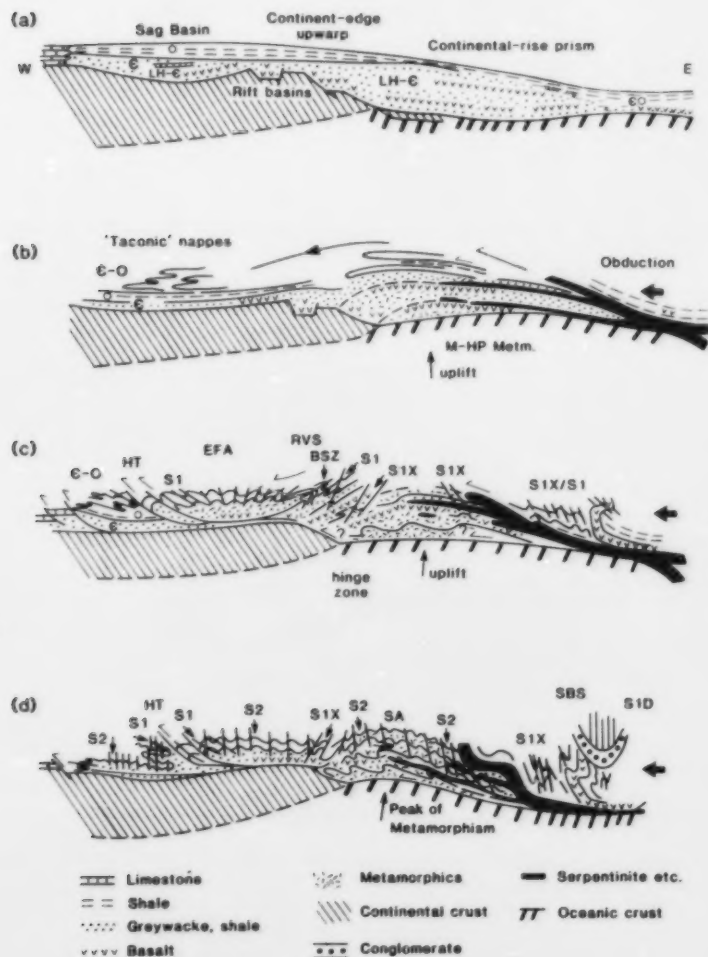


Figure 45. Sketch cross sections illustrating deformational history. **a)** Paleogeographic reconstruction of rifted continent margin and continental-rise prism—Early Ordovician. **b)** Obduction of ocean floor over continental-rise prism; thrusting, overthickening, metamorphic doming (medium to high pressure) and nappe emplacement—early Middle Ordovician. **c)** Continuing uplift, development of bedding schistosity with gravity sliding, thrusting and early folding, followed by cross folding in the schist belt and Eastern Zone—Middle Ordovician to Early Silurian. **d)** Tightening of structures, development of the Sutton Anticline and late superposed cleavage accompanied by peak of thermal metamorphism—Silurian (?) to Middle Devonian. In the east, folding of Siluro-Devonian rocks formed the Sargent Bay Syncline (with axial-plane cleavage S_{1D}) in the Middle Devonian Acadian Orogeny. EFA = Enosburg Falls Anticlinorium; RVS = Richford-Valcourt Syncline; SA = Sutton Anticline; HT = Hinesburg Thrust; BSZ = Brome Slide Zone; SBS = Sargent Bay Syncline.

of folds with axial-plane slaty cleavage developed to form the Enosburg Falls Anticlinorium (Fig. 45c). The Hinesburg fault has been interpreted to be part of an extensive thrust system rooting downward (Stanley and Ratcliffe, 1985), and in the case of the Champlain Thrust, to the west, cutting and carrying the Precambrian basement of the Green Mountains more than 80 km as an allochthonous sheet (Rowley, 1982). In the Sutton area, the Hinesburg Thrust has only a small movement and the fanning axial planes of the Enosburg Falls Anticlinorium and the associated eastward thinning of the Oak Hill strata into the Mansville facies suggests a concave upward form for the Hinesburg Thrust (Fig. 45c). These structures are taken to imply a gravity-slide mechanism, as the Oak Hill rocks (in the Hinesburg slice of Stanley and Ratcliffe, 1985) slid westward off the rising metamorphic core zone situated in the continental-rise prism. The boundary between these units, the Brome Slide Zone, is interpreted to be a complex strain discontinuity located above the crustal hinge zone that during its subsequent development became folded by the Richford-Valcourt Syncline (Fig. 45c). The metamorphic boundary of the albite porphyroblasts zone coincides with the Brome Slide Zone in the west and the serpentinite belt in the east (Fig. 23) suggesting that, although the peak of metamorphism was later, the developing metamorphic welt controlled the location of structural breaks—a situation similar to other metamorphic core complexes (Coney, 1980).

Differences in structure between this area and those to the north and south may be attributed to the original paleogeography of the Quebec re-entrant in the Precambrian Shield (Dewey and Kidd, 1974; Doolan et al., 1982; Stanley and Ratcliffe, 1985); nappe development to the north and south appears to be more intensive with shingle-like overlaps where compressive impingement against basement salients occurred (Stanley and Ratcliffe, 1985) (Fig. 46b).

In the continental-rise prism itself, uplift continued possibly associated with fault-bend folding of deep thrusts (St. Julien et al., 1983; Stanley and Ratcliffe, 1985). Longitudinal shear on the bedding schistosity caused minor east-west reclined folds to develop across the regional trend (Fig. 46c). These extended upward to impinge on the lower part of the Oak Hill sequence in the Richford-Valcourt Syncline. The shearing movements were probably caused by differential lateral adjustments as the collapsing continental-rise prism was bent into the re-entrant in the Precambrian Shield (Fig. 22 and 46b,c); it is possible that the movements were enhanced by flow off rising metamorphic highs developed along the zone. Although northerly-trending folds occur in the eastern oceanic zone, a series of reclined to vertical folds predominate and they probably owe their spectacular development to their location in the concave core of the re-entrant (Fig. 46c). Movement into the re-entrant probably also caused the shredding out of the ophiolite belt to the south (Doolan et al., 1982) (Fig. 46c).

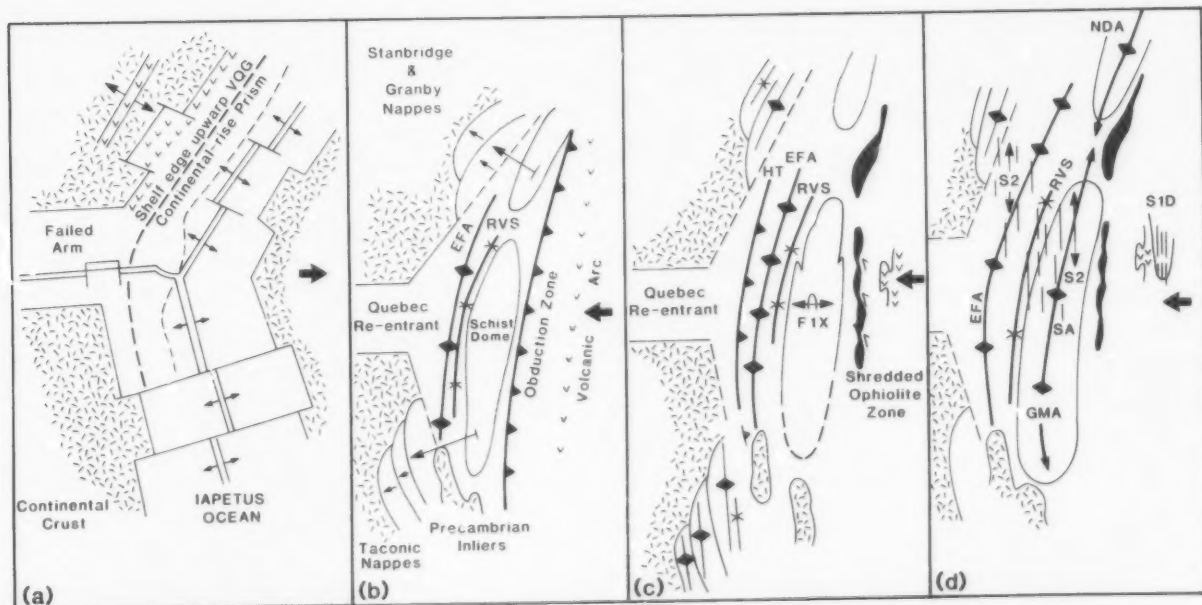


Figure 46. Sketch maps illustrating deformation history: **a)** Development of rift ocean and failed arm to produce Quebec re-entrant—late Precambrian to Cambrian. **b)** Obduction collision of the volcanic arc with strong thrusting and nappe formation at salients; development of early upright folds—early Middle Ordovician. **c)** Bending into Quebec re-entrant develops reclined cross folds and steep folds, and disruption of the ophiolite belt—Late Ordovician. **d)** Final development of the Sutton Anticline, transecting late cleavage and peak metamorphism—Silurian (?) or Middle Devonian. EFA = Enosburg Falls Anticlinorium; RVS = Richford-Valcourt Syncline; SA = Sutton Anticline; NDA = Notre Dame Anticline; GMA = Green Mountain Anticlinorium; VQG = Vermont-Quebec Geanticline; HT = Hinesburg Thrust.

The Taconic Orogen thus formed by subduction-obduction processes associated with the initial collision of an island arc (Osberg, 1978; Williams, 1984) was quickly eroded and overlapped by shale-carbonate sequences in the Late Silurian to Early Devonian as the focus of orogenic activity shifted eastward (Poole, 1976; Bradley, 1983). Final closure of the Iapetus Ocean occurred diachronously from Early to Late Devonian times as a continent-continent collision, and this Acadian Orogeny was marked by crustal thickening and extensive granite intrusion (Osberg, 1978; Schenk, 1978; Taylor and Toksöz, 1982; Williams, 1984). Locally the whole process from sedimentation to metamorphism and granite intrusion took only 30 Ma (Naylor, 1971). Mild folding of the sediments covering the Taconic Orogen was probably accommodated by flattening and shearing along pre-existing foliations in the basement, and this yielded the marked structural discordance between the Sargent Bay Syncline and the basement folds in the Eastern Zone (Fig. 45d and 46d).

In the Central and Western zones, west of the ophiolite suture line, deformation was terminated by a tightening phase in which the Sutton Anticline took on its final form. Thermal metamorphism reached its peak during and after this deformation so that albite porphyroblasts overgrew late crenulations. The associated late crenulation cleavage was slightly oblique to the earlier structures, and extended beyond the Sutton Anticline obliquely across the folds of the Enosburg Falls Anticlinorium (Fig. 46d). This deformation may represent the termination of the Taconic Orogeny or more likely a reactivation effect during the Middle Devonian Acadian Orogeny. Steep reverse faults developed about this time and they are concentrated where stress perturbations developed, especially at the carbonate-bank scarp (Rodgers, 1968).

Finally, after a long period of erosion, the complex structures still exert a profound influence on the present-day topography of the region.

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APPENDIX

Age determinations in the Sutton area

Rock	Locality	Published Ref. Nos.	Age Ma $\pm$	Type	Reference
1 Pinnacle greywacke	West Sutton	GSC61-182	427 30	K-Ar Musc.	Lowden et al., 1963
2 Pinnacle greywacke	West Sutton	PB78-66	427 11	K-Ar Musc. ¹	Poole (pers. comm.)
			390 13	K-Ar Musc.	
3 Pinnacle greywacke	Sutton Junction	-	411 5.7	K-Ar Musc.	Knight, 1976
4 Gilman phyllite	S of East Dunham	-	388 5.4	K-Ar Seric.	Knight, 1976
5 West Sutton slate	West Sutton	-	382 5.3	K-Ar WR	Knight, 1976
6 Mohawk Lake fm	SE of Sutton	GSC61-183	448 30	K-Ar Musc. ²	Lowden et al., 1963
7 Sallys Pond fm	Glen Sutton	-	421 5.8	K-Ar Musc.	Knight, 1976
			372 5.2	K-Ar Biot.	Knight, 1976
8 Bolton Glen fm	SE of Knowlton	GSC78-200	367 12	K-Ar Musc.	Wanless et al., 1979
			411 4	Rb-Sr WR/Musc. ³	Nieuwlands (pers. comm.)
9 Sallys Pond fm	SE of Knowlton	GSC78-201	380 13	K-Ar Musc. ⁴	Wanless et al., 1979
10 St. Daniel Fm	Vale Perkins	56	365 ?	K-Ar WR	Harper, 1968

Musc. = muscovite; Seric = sericite; Biot. = biotite; WR = whole rock

Samples are located on Figure 23. All dates have been corrected to current standards (Steiger and Jager, 1977)

1. This sample was collected anew and redetermined in 1979 to check for possible errors in the 1963 measurements reported by Wanless et al. (1973). The result is the same as the 1963 analysis, but the other result from a repeat analysis on a different sample gives a younger date. This greywacke has a well-developed slaty-type cleavage with no sign of late F_2 effects. The texture is illustrated in Figure 24.
2. Although Wanless et al. (1979) cast doubt on this result because younger ages were obtained for Sutton schists in later analyses (see comments 3 and 4) the redetermination mentioned in comment 1 also confirms the likely validity of this result. In the specimen, straight micas define the S_{1X} schistosity and lie in the S_2 cleavage direction; albite porphyroblasts enclose F_2 microfolds (Rickard, 1965).
3. This specimen was chosen because it showed a strong development of early fold (F_{1X} , S_{1X}) structures, so that even though the recrystallization appears to be post- F_2 , the possibility of an older hybrid date was

enhanced. The texture is similar to that illustrated in Figure 29a,b. Since the age is similar to that in specimen GSC78-201 (see comment 4), recrystallization and degassing during F_2 was probably complete. A Rb-Sr age was also determined for this specimen by analyzing separately the whole rock, separated muscovite and the whole rock minus muscovite. This gave a model 1 age of 411 ± 4 Ma. Analysis by R. Nieuwlands at the Australian National University.

4. This specimen was chosen to contrast with specimen GSC78-200. Only a trace of the early F_{1X} folding remains, straight recrystallized muscovites define F_2 microfolds, the S_2 crenulation cleavage is well developed and near euhedral albite porphyroblasts that contain F_2 microfolds as inclusions have prism faces oriented in the S_2 axial plane. Textures are similar to those illustrated in Figures 31a,b and 37a,b. Recrystallization during and after F_2 appears to have been complete so that total degassing and resetting of micas can be confidently expected.

**MAPS
NOT
FILMED**

**CARTES
OMISES**